Bulletin

of

The Geological Society of America

VOLUME 65 February 1954 NUMBER 2

CONTENTS

	Pages
Criteria for the mode of emplacement of the alkaline stock at Mount Monadnock, Vermont. By Randolph W. Chapman	97–114
Genesis of Precambrian granitic pegmatites in the Denver Mountain Parks area, Colorado. By Margaret Fuller Boos	115–142
Indonesian archipelago: a geophysical study. By F. A. Vening Meinesz	143–164
Gravity anomalies and structure of the West Indies, Part I. By Maurice Ewing and J. Lamar Worzel	165–174
Geology of alluvial fans in semiarid regions. By Erich Blissenbach	175–190
Short Notes	
Deep-sea channels and delta of the Monterey submarine canyon. By Robert F. Dill, Robert S. Dietz, and Harris Stewart	191–194
Gravity anomalies and structure of the West Indies, Part II. By J. Lamar Worzel and Maurice Ewing	195–200

Subscription $15.00 per year

Printing Office: Mt. Royal & Guilford Aves., Baltimore 2, Md.

Communications for publications should be addressed to The Geological Society of America, Dr. H. R. Aldrich, Secretary, 419 West 117 Street, New York, N. Y.

Please give notice of change of address 4 weeks in advance. Claims for non-receipt of the Bulletin of a given month must be sent to the Secretary of the Society before the end of the second succeeding month.

Entered as second-class matter at the Post-Office at Baltimore, Md., under the Act of Congress of July 16, 1894.

Accepted for mailing at special rate of postage provided for in Section 1103 Act of October 3, 1917, authorized on July 3, 1918.

PAPERS IN PRESS FOR FORTHCOMING ISSUES

GEOPHYSICAL INVESTIGATIONS IN THE EMERGED AND SUBMERGED ATLANTIC OCEAN PLAIN, PART VIII: GRAND BANKS AND ADJACENT SHELVES. By Frank Press and Walter Beckmann

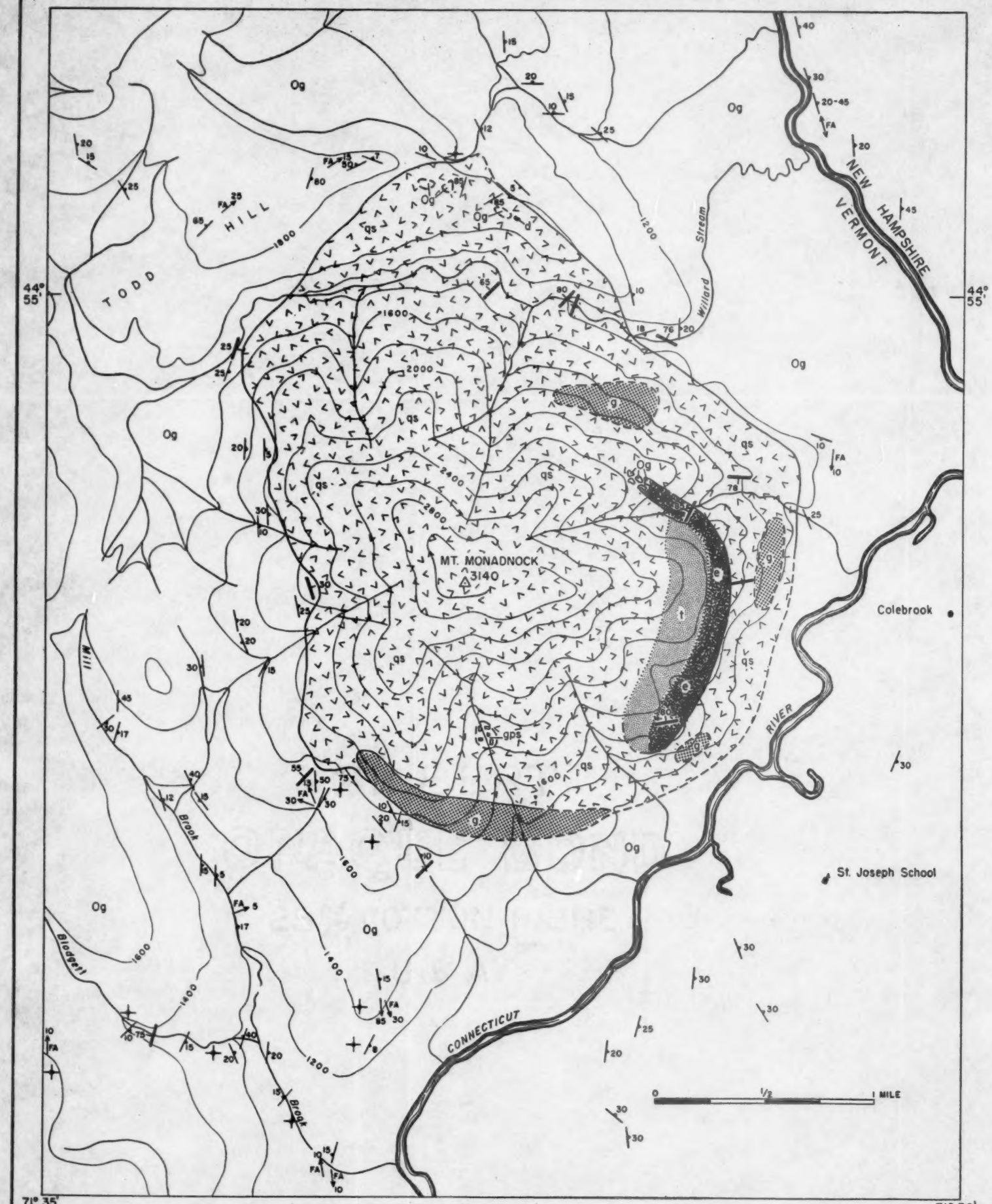
CORRELATION OF THE ORDOVICIAN FORMATIONS OF NORTH AMERICA. By Ordovician Subcommittee, W. H. Twenhofel, *Chairman*

CORRELATION OF THE CRETACEOUS FORMATIONS OF GREENLAND AND ALASKA. By Ralph W. Imlay and John B. Reeside, Jr.

BEAR RIVER AND ASPEN FORMATIONS IN WILLOW CREEK, TETON COUNTY, WYOMING. By A. LaRocque and C. D. Edwards

GEOLOGIC PROBLEMS IN THE SOUTHEAST MISSOURI LEAD DISTRICT. By E. H. Ohle and John S. Brown *Editors*

SECRETARY'S NOTE: *Numerical Structure Factor Tables*, By M. J. Buerger (GSA Special Paper 33) has been reprinted and is available at $1.50 per copy (Cloth bound). Remittance should accompany order.



LEGEND

IGNEOUS AND METAMORPHIC ROCKS

MISSISSIPPIAN (?) — White Mountain magma series

- Alkaline dike, dipping
- Alkaline dike, vertical
- g — Granite
- qs — Quartz syenite
- gps — Gray porphyritic syenite
- t — Transition rock
- d — Diorite
- Q — Essexite

ORDOVICIAN

- Og — Gile Mountain formation (?) (Schist and quartzite)

CONTACTS

- Accurate
- Approximate and inferred
- Gradational

STRUCTURE SYMBOLS

- Strike and dip of schistosity
- Horizontal schistosity
- Plunge of fold axis
- Horizontal fold axis

GEOLOGIC MAP OF MOUNT MONADNOCK, VERMONT

Topography based on Averill, Vermont, quadrangle of U. S. Geological Survey

CH

T
studi
of en
T
quar
and
arcu
Late
T
Mou
and
dike
The
dike
dike
coun
F
acco

Intr
N
P
Ge
Pe
C
F
U
T
C
C
C
Str
C
N
F
C
S
T
F
F
T
Str
En
Re

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 65, PP. 97-114, 2 PLS., 2 FIGS. FEBRUARY 1954

CRITERIA FOR THE MODE OF EMPLACEMENT OF THE ALKALINE STOCK AT MOUNT MONADNOCK, VERMONT

By Randolph W. Chapman

Abstract

The alkaline stock at Mount Monadnock, Vermont, described briefly by Wolff (1929), has been restudied in detail. Its petrography and structure are discussed here and conclusions are drawn as to its mode of emplacement.

The stock consists of plutonic and hypabyssal rocks which intrude folded Ordovician (?) schist and quartzite. The longer axis, trending north-northwest across the strike of the country rock, is 3 miles long, and the shorter one about 2½ miles. Essentially the stock consists of quartz syenite, but it encloses a long arcuate mass of older essexite and transition rock, and along its eastern edge are later intrusions of granite. Late dikes of various compositions cut the plutonics and surrounding metamorphic rocks.

The following facts have been established: (1) The igneous rocks are typical representatives of the White Mountain magma series (Mississippian?), quite lacking in foliation and lineation. (2) The stock is discordant and has an elliptical ground plan. (3) In detail the boundary is irregular and characterized by abundant dikes and xenoliths. (4) The igneous rocks make sharp contacts with the metamorphosed country rock. (5) The arcuate mass of older essexite is undoubtedly a screen. (6) The small bodies of late granite resemble ring dikes. (7) The stock is cut by prominent sets of steeply dipping radial and tangential joints. (8) The late dikes show radial and tangential patterns. (9) Along the northern and southern margins of the stock the country rock shows strikes and dips which differ from the regional ones.

From these criteria it is concluded that the plutonic rocks have invaded the crust by cauldron subsidence accompanied by the stoping of large arcuate slabs and smaller blocks from the walls of the magma reservoir

CONTENTS

TEXT

	Page
Introduction	98
Nature of the project	98
Physiography	98
General geologic relations	99
Petrography of the intrusive rocks	99
General statement	99
Essexite	99
Diorite	101
Transition rock	102
Gray porphyritic syenite	102
Quartz syenite	103
Granite	104
Structure of the stock	105
General statement	105
Main quartz syenite body	105
Intrusion of gray porphyritic syenite	106
Granitic intrusions	106
Screen	106
Transition zone	107
Dikes	107
Boundary of the stock	107
Joints	110
Structure of the country rock	110
Emplacement of the stock	111
References cited	114

ILLUSTRATIONS

Figure	Page
1. Map of the joints in the quartz syenite stock at Mount Monadnock, Vermont	109
2. Diagrammatic sketches illustrating the suggested mode of emplacement of the alkaline stock at Mount Monadnock, Vermont	113

Plate	Facing page
1. Geologic map of Mount Monadnock, Vermont	97
2. Mount Monadnock, Vermont	106

TABLES

Table	Page
1. Estimated modes of the essexite in volume per cent	100
2. Chemical compositions of essexite, quartz syenite, and hornblende (hastingsite) from Mount Monadnock, Vermont	100
3. Estimated modes of the diorite in volume per cent	102
4. Estimated modes of the transition rock in volume per cent	102
5. Estimated modes of the gray porphyritic syenite in volume per cent	103
6. Estimated modes of the quartz syenite in volume per cent	104
7. Estimated modes of the granite in volume per cent	105

Introduction

Nature of the Project

Mount Monadnock, Vermont, is located on the west side of the Connecticut River opposite the village of Colebrook, New Hampshire. It is a stock composed of resistant rocks of the White Mountain magma series (Mississippian?), an alkalic suite, representatives of which are abundantly distributed throughout north-central New England (Greenwood, 1951, Pl. 4). In eastern Massachusetts and Rhode Island are closely related alkalic intrusions believed to be of about the same age. To the northwest, in the Province of Quebec, similar intrusive rocks form stocklike bodies underlying the Monteregian Hills. Mount Monadnock, the most northerly rampart of the White Mountain magma series in New England, thus occupies a strategic position between this magma series and the rocks of the Monteregian Hills.

Wolff (1929) mapped Mount Monadnock, described the geology briefly, and presented optical and chemical data on some of the minerals and rocks. However, he prepared his own topographic map, and, since it was erroneous in many respects, he was unable to plot the distribution of rock types accurately. Moreover, several rock types were overlooked in the field, and few critical data were obtained on the structural features of the intrusion and the surrounding rocks.

In view of the strategic position of the Mount Monadnock stock and the rather inadequate geologic data available on it, the writer undertook a detailed study of it in the summer of 1949. His purpose was to prepare an accurate geological map of the pluton and to attempt to deduce its mode of emplacement from its structural and petrologic features. About 5 weeks was spent in field mapping. Reconnaissance mapping was done in a zone 4 or 5 miles wide around the mountain in order to ascertain the regional structure of the metamorphic rocks. Much of this outer zone lies outside the area of Plate 1.

The project was made possible by a generous grant in aid from the Research Fund of the Scientific Research Society of America which financed the field expenses and the preparation of thin sections. The Geological Society of America kindly paid for the final drafting of the maps and diagrams. Kenneth N. Weaver ably assisted the writer in the field, and Miner B. Long and John C. Whitaker helped in the petrographic studies. The writer is pleased to acknowledge his deep gratitude for all this assistance.

The first part of this paper describes the petrography of the intrusive rocks of Mount Monadnock in an attempt to: (1) give the reader a clearer understanding of the second part; (2) help in correlating these rocks of the White Mountain magma series with those in other areas. The second part deals with the structure of the stock and attempts to explain its mode of emplacement by using certain available criteria.

Physiography

Mount Monadnock, Vermont, called Monadnock Mountain on the Averill quadrangle of the United States Geological Survey, was named by the early settlers after the better-known mountain of physiographic fame in southwestern New Hampshire. It is a broad conical peak, rising boldly above the surrounding country (Pl. 2, fig. 1). Its major portion is in the township of Lemington, but its northeast flank extends into the township of Canaan. In plan and profile the mountain simulates other New England mountains of like geologic structure, such as Cherry Mountain in New Hampshire, Pleasant Mountain in Maine, and Ascutney Mountain in Vermont. Its summit attains an elevation of 3140 feet and the Connecticut River at its base is at an elevation of 1000 feet, making a maximum relief of 2140 feet.

Mount Monadnock is the terminus of a long, rugged chain of mountains extending northwestward across northern New Hampshire, but it is physically separated from this chain by the valley of the Connecticut River. For many miles north and west of the mountain the topography is less rugged and is characterized by innumerable hills with extensive intervening swamps.

Except for a flat bench on its southern margin (Pl. 2), the mountain is a symmecrical cone the sides of which slope uniformly from the central summit. On these slopes the many active streams draining the mountain have de-

veloped a radial pattern. Willard Stream on the north side, and Mill Brook on the south, are the two main streams that empty into the Connecticut River. Willard Stream has intrenched itself firmly into the north flank of the mountain, forming a deep and scenic ravine with Todd Hill to the north.

Thick deposits of glacial outwash mantle much of the lower slopes of Mount Monadnock, especially along the east and southeast sides where they are augmented by river alluvium, and on the west side where swamps are widespread. Glacial debris extends far up the north and northwest slopes, but on the south and southeast sides it is generally absent above an elevation of 1200 feet, and outcrops of bedrock are numerous.

Except for limited fields and pastures at the foot of the east and south slopes, the mountain is densely forested. A fire observation tower on the summit of the mountain may be reached by a good trail which begins 100 yards southwest of the bridge over the Connecticut River west of Colebrook.

General Geologic Relations

Mount Monadnock is carved from an elliptical stock of igneous rocks of the White Mountain magma series (Mississippian?) intruded into folded schist and quartzite which strike north-northeasterly (Pl. 1). Its longer axis, trending north-northwest across the strike of the country rock, is 3 miles long, and the shorter one about 2½ miles. The boundary of the stock, although irregular in detail, has been located accurately, except on the southeast side of the mountain where it is buried by river and glacial deposits.

A small hill due east of St. Joseph School, 2 miles southeast of Colebrook, is underlain by syenite resembling that on Mount Monadnock, a fact that led Wolff to conclude that the stock may extend southeastward across the Connecticut River. Such a conclusion is quite logical, but it is equally possible that the syenite east of St. Joseph School may belong to a different intrusion. Until the geology of that mountainous area to the southeast has been studied, it is perhaps better to assume that the southeast margin of the Mount Monadnock stock lies wholly on the northwest side of the Connecticut River.

The Mount Monadnock stock is actually composite and consists of successive plutonic intrusions followed by a retinue of dikes. From oldest to youngest the igneous rocks are: essexite; diorite (as inclusions); transition rock; gray porphyritic syenite; quartz syenite (dark fine-grained type, light-colored fine-grained type, medium-grained type); granite; dike rocks.

This sequence was determined by observing contact features such as apophyses (Pl. 2, fig. 2), inclusions, and chilled borders. The order is believed to be correct with two possible exceptions: (1) The relative ages of the essexite and diorite are unknown, since the two rocks have not been found in contact, but analogy with occurrences of these rocks in other areas suggest that the essexite is older; (2) most of the dikes do not cut the granite, but intrude the medium-grained quartz syenite and older rocks. However, analogy with other similar areas indicates that the dikes are the youngest rocks. The relative ages of the many dike types are not known.

Petrography of the Intrusive Rocks

General Statement

The petrography of all the intrusive rocks of Mount Monadnock, except the dike rocks, is described in this section. Although the present paper deals mainly with structure, petrographic descriptions are included for two reasons: (1) to assist in understanding the origin and mode of emplacement of these rocks, and (2) to help to correlate these rocks of the White Mountain magma series with those in other areas. The reader may obtain more information about the optical properties and chemical compositions of the rock-forming minerals in Wolff's (1929) report.

It is not essential to read the detailed petrography in this section in order to understand the structural part of the paper, because the megascopic character of each rock type is mentioned briefly in the section on Structure of the Stock.

Essexite

This rock is similar to the essexite of Cuttingsville, Vermont, as described by Eggleston

(1918). Megascopically, it is medium-grained and even granular and varies from medium to dark gray. The dark minerals compose about 50 per cent of the rock. Along the eastern contact with the syenite the grain size is generally about 1 mm, but elsewhere it ranges from 1 mm to 4 mm. Several specimens are subporphyritic with phenocrysts of plagioclase and femags. Scattered pyrite can be seen in a few specimens. The rock is essentially without structure; no banding or flow structure are visible. The feldspars weather chalky white, whereas the dark minerals become stained with deep reddish-brown iron oxide.

Microscopically the essexite is allotriomorphic and even-grained to seriate. One specimen is slightly porphyritic. In general the minerals are notably irregular, and in some cases hornblende, pyroxene, and biotite are strongly poikilitic with inclusions of apatite and magnetite. The essential minerals are plagioclase, hornblende, biotite, pyroxene, and potash feldspar, although the latter is abundant in only one section. In three slides pyroxene composes between 5 and 20 per cent of the rock. Accessory minerals include apatite, magnetite, sphene, and zircon. In one slide a few grains of quartz were detected. Secondary minerals such as kaolinite, sericite, and chlorite are present in varying amounts. The essential minerals range in grain size from 0.25 mm to 3 mm and average about 1 mm. Table 1 shows estimated modes of four thin sections.

A specimen of essexite was analyzed chemically by Wolff (1929, p. 13), and its composition is shown in Table 2.

Andesine-labradorite (Ab_{50}) makes up about half the rock. It occurs either as anhedra 0.2 mm to 2.5 mm in diameter or as tabular laths 0.1 mm to 2 mm long. It is generally untwinned and in many places is progressively zoned. Many of the larger crystals show cores which are twinned polysynthetically and which grade outward into untwinned rims. Many are altered to kaolinite or sericite.

Potash feldspar is abundant in only one section examined; it forms anhedra which resemble the plagioclase. In another slide only a few grains were observed. It is because of the presence of this mineral that the rock is called essexite.

TABLE 1.—ESTIMATED MODES OF THE ESSEXITE IN VOLUME PER CENT

Specimen No.	MM 6	MM 92	MM 103	MM 113
Andesine-labradorite	55	45	40	50
Potash feldspar	15	...	...	tr.
Hornblende	10	25	25	25
Biotite	13	20	7	10
Pyroxene	...	5	20	10
Accessories	7	5	8	5
Total	100	100	100	100

TABLE 2.—CHEMICAL COMPOSITIONS OF ESSEXITE, QUARTZ SYENITE, AND HORNBLENDE (HASTINGSITE) FROM MOUNT MONADNOCK, VERMONT

	1	2	3
SiO_2	46.09	69.43	38.72
TiO_2	4.49	.18	5.03
Al_2O_3	17.29	15.67	10.94
Fe_2O_3	1.63	.36	4.33
FeO	7.26	2.50	26.03
MgO	8.04	.12	2.19
CaO	7.89	1.13	8.55
Na_2O	3.68	5.11	1.39
K_2O	1.92	4.97	1.25
H_2O+	.77	.42	1.22
H_2O-	.12	.01	.....
ZrO_2	n.d.	.06	.....
P_2O_5	.90	.06	.....
S	.10	.07	.....
MnO	.16	.06	.88
BaO	.09	none	.....
SrO	tr.	n.d.	.....
Total	100.43	100.15	100.53

1. Essexite from Mount Monadnock, Vermont (Wolff, 1929, p. 13).
2. Quartz syenite from Mount Monadnock, Vermont, (Wolff, 1929, p. 13).
3. Hornblende (hastingsite) from quartz syenite from Mount Monadnock, Vermont (Wolff, 1929, p. 9).

Brown biotite, composing 10 to 22 per cent of the rock, is pleochroic from light yellow to deep red brown. It occurs in large sheets, up to 3.5 mm in diameter, and in many cases encloses magnetite and apatite poikilitically. Some biotite forms rims on magnetite included in pyroxene, suggesting late growth after the rock solidified.

Most of the hornblende is a light-green pleochroic variety, but in one section it is deep red brown. It is commonly associated with biotite in irregular interstitial aggregates but also occurs as large-zoned phenocrysts with dark-green centers and lighter-green rims, or as reaction rims on pyroxene. The hornblende composes 5 to 25 per cent of the rock and crystals range up to 3 mm in diameter.

In one of the slides examined pyroxene is absent, but in the others it makes up 5 to 20 per cent of the rock. It is generally a light-green nonpleochroic variety occurring in irregular aggregates. In one slide it occurs as large brown phenocrysts, 3 mm long, surrounded by thick green reaction rims of hornblende. The pyroxene from specimen MM 103 (Table 1) from near the center of the essexite body has the following optical properties: $\alpha = 1.690$, $\beta =$ ca. 1.695, $\gamma = 1.717$; optically positive; $2V =$ ca. 60°; Y = b, $Z \wedge c = 42°$; pale smoky pink in section; nonpleochroic. Apparently this mineral belongs to the diopside-hedenbergite series and consists of 65 per cent diopside and 35 per cent hedenbergite (Winchell, 1951, Fig. 302). The optical properties correspond closely to those of the diopside from the gabbro of Tripyramid Mountain, New Hampshire (Chapman and Williams, 1935, p. 513–514).

Magnetite is an abundant accessory in all slides examined and in one slide totals 5 per cent. It is usually anhedral but occasionally is subhedral to anhedral. Apatite is also an abundant accessory ranging in length from less than 0.05 mm to 1.5 mm. Sphene was detected in only two slides, although in one of these it is extremely abundant. It is generally subhedral to anhedral and ranges from 0.2 mm to 1.5 mm across. In two slides zircon was found in minute amounts.

In the essexite, apatite, magnetite, zircon, and sphene were the first to crystallize: they are enclosed by all other minerals and are occasionally euhedral. Pyroxene, plagioclase, and potash feldspar crystallized in anhedra and appear later in the sequence. Hornblende developed reaction rims on the earlier-formed pyroxene, while biotite grew on magnetite crystals included in the pyroxene. Biotite shows a marked tendency to include most of the earlier-formed minerals poikilitically, suggesting growth in a solid medium. The paragenesis is thus deduced as: apatite, magnetite, zircon, sphene; pyroxene, plagioclase, potash feldspar; hornblende, biotite; quartz.

Diorite

Diorite inclusions, believed to be remnants of an earlier intrusion, are abundant in the quartz syenite stock. The diorite is a fine- to medium-grained, even granular to subporphyritic, medium- to very dark-gray rock. A few specimens suggest a hornfels texture. The phenocrysts are femags averaging about 3 mm in length and reaching a maximum of 1 cm. Clusters of quartz and feldspar scattered through the diorite range in diameter from several millimeters to several centimeters. The diorite is essentially structureless.

Microscopically, the diorite is usually allotriomorphic and even-grained, but occasionally is vaguely subporphyritic. Within an individual thin section, however, the grain size is constant. The average grain size of the specimens studied ranges from 0.08 to 0.2 mm. Generally the dark minerals are equigranular with irregular borders and are evenly disseminated through the rock. An occasional phenocryst of hornblende up to 2.5 mm long was observed. The essential minerals are oligoclase, hornblende, biotite, and pyroxene. In two sections quartz occurs in clusters associated with oligoclase. Here the quartz and oligoclase are anhedral and equigranular and somewhat larger than the groundmass. The average size of the grains in the clusters is about 1 mm. One cluster measures 1.5 cm across.

The oligoclase (Ab_{70-75}) in the smaller grains is equigranular, while the somewhat larger oligoclase occurs in subhedral tabular laths 0.08 to 0.7 mm in length. Pyroxene, hornblende, and biotite are optically similar to those in the essexite. In one slide the pyroxene was surrounded by thick reaction rims of green hornblende with biotite penetrating the rims.

Accessories include apatite, magnetite, sphene, and zircon. While apatite and magnetite are abundant, sphene is not nearly so abundant as in the essexite. The evenly disseminated character of the hornblende, biotite, and pyroxene suggests a hornfels texture, but it is poorly developed. The paragenesis of the minerals is

identical with that determined for the essexite. The estimated modes of three thin sections are given in Table 3.

Transition Rock

The transition rock is light gray, medium-grained, and even granular. It differs from the essexite mainly in its lower content of dark minerals which constitute 15 to 25 per cent. Quartz is present in small amounts. The feldspar weathers chalky white and in some specimens is stained with iron oxide.

TABLE 3.—ESTIMATED MODES OF THE DIORITE IN VOLUME PER CENT

Specimen No.	MM 14	MM 29	MM 98
Oligoclase	65	75	55
Hornblende	7	3	15
Biotite	12	7	15
Pyroxene	5	10	.....
Quartz	5	.....	10
Accessories	6	5	5
Total	100	100	100

Under the microscope the transition rock is allotriomorphic and seriate with a general absence of the poikilitic tendencies of the essexite. The essential minerals are andesine, potash feldspar, quartz, hornblende, and biotite. Andesine occurs in tabular laths, usually twinned polysynthetically, and in many slides is progressively zones. The index of refraction suggests approximately the same chemical composition as for the andesine of the essexite. The potash feldspar is untwinned and is interstitial. The hornblende is somewhat darker than that described in the essexite, whereas the biotite is identical in optical properties with that in the essexite. Hornblende, biotite, magnetite, and sphene occur in irregular clusters. Quartz is interstitial. Accessory minerals include magnetite, sphene, and apatite. Sphene and magnetite are abundant and may compose as much as 5 per cent of a specimen. The feldspars are highly altered to kaolinite and some sericite. The estimated modes of two thin sections are given in Table 4.

A comparison of the transition rock and the quartz syenite (described later) reveals the following differences: (1) The transition rock tends to be seriate, whereas the quartz syenite is mostly equigranular. (2) There is no microperthite in the transition rock, but more plagioclase in the transition rock than in the quartz syenite. (3) The plagioclase of the transition rock (Ab_{60}) is less sodic than that of the quartz syenite (Ab_{70}). (4) The transition rock is richer in dark minerals than the quartz syenite. (5) The biotite of the transition rock is pale yellow to deep red brown, whereas that of the quartz syenite is yellow green to olive brown. (6) Allanite and zircon are present in the quartz syenite.

TABLE 4.—ESTIMATED MODES OF THE TRANSITION ROCK IN VOLUME PER CENT

Specimen No.	MM 94	MM 100
Andesine	46	44
Potash feldspar	25	20
Quartz	5	5
Hornblende	10	15
Biotite	10	10
Magnetite	2	2
Sphene	1	3
Accessories	1	1
Total	100	100

It is believed that the transition rock is a hybrid which represents quartz syenite contaminated with essexite.

Gray Porphyritic Syenite

The gray porphyritic syenite is a dark-gray, structureless rock with conspicuous phenocrysts of biotite and feldspar in a fine-grained groundmass. Feldspar phenocrysts range from 2 to 4.5 mm, and biotite phenocrysts from 1.5 to 2.5 mm. A conspicuous feature is the presence of angular inclusions, averaging about 1.5 inches across, composed of about equal amounts of dark and light minerals. Within the inclusions are darker clots of pure biotite and rosettes of radiating actinolite surrounded by narrow rims of biotite.

Two thin sections of gray porphyritic syenite were studied, and their estimated modes are

listed in Table 5. The essential minerals are microperthite, potash feldspar, oligoclase-andesine (Ab_{70}), quartz, biotite, hornblende, and pyroxene. One section contains 4 per cent pyroxene, and another less than 1 per cent, occasionally with reaction rims of hornblende.

TABLE 5.—ESTIMATED MODES OF THE GRAY PORPHYRITIC SYENITE IN VOLUME PER CENT

Specimen No.	MM 57	MM 58
Microperthite	70	79
Potash feldspar	10	5
Plagioclase	10	5
Quartz	5	2
Biotite	2	5
Hornblende	3	.....
Pyroxene	.....	4
Total	100	100

Accessory minerals usually amount to 1 per cent or less and have been omitted from the above modes.

Several feldspar phenocrysts consist of potash feldspar cores with borders of microperthite. Accessory minerals include magnetite, sphene, apatite, and allanite. The gray porphyritic syenite is conspicuously porphyritic with large, slightly rounded phenocrysts in a fine-grained groundmass averaging 0.2 mm in grain size.

The gray porphyritic syenite differs from the quartz syenite (described later) in being dark gray, having slightly more pyroxene, less quartz, larger and more rounded phenocrysts, a finer groundmass, and zoned feldspars.

Quartz Syenite

In hand specimen the quartz syenite is typically structureless, tough, light gray to pink, and usually even-grained to subporphyritic. It is fine- to medium-grained with conspicuous feldspar laths and fine-grained, equidimensional dark minerals. A few specimens are distinctly porphyritic. The feldspar laths range from 0.05 mm to 5 mm long and average about 1 mm in width. Less commonly, however, they are equidimensional. The dark minerals range from a fraction of a millimeter to 2 mm. Most typical specimens are light gray and fine-grained, with about 8 per cent dark minerals. The rock weathers with a typically blotchy, reddish-brown iron-oxide stain.

Conspicuous features of the quartz syenite are dark-gray diorite inclusions composed of about equal amounts of light and dark minerals. These average 1.5 inches in cross section.

Under the microscope the quartz syenite is hypidiomorphic granular and consists essentially of microperthite, quartz, biotite, hornblende, potash feldspar, and plagioclase. In general two textures prevail. One type is predominantly hypidiomorphic, more or less equigranular, and fine-grained. The other is hypidiomorphic, somewhat finer-grained, and subporphyritic to porphyritic with some tendency to a glomeroporphyritic texture. The phenocrysts are predominantly microperthite. One specimen is intergranular with lathlike feldspars showing a crude trachytoid arrangement. The grain size for all specimens ranges from 0.03 mm to 2 mm and averages 0.8 mm.

The estimated modes listed in Table 6 show that the quartz content ranges between 5 and 20 per cent. Of the 17 specimens studied, 4 show more than 15 per cent quartz and are therefore strictly granites rather than quartz syenites.

Accessory minerals are magnetite, sphene, apatite, zircon, pyroxene, and allanite, the latter pleochroic in shades of reddish brown. Chlorite and hematite are secondary.

The chemical composition of a specimen of quartz syenite reported by Wolff (1929, p. 13) is shown in Table 2.

The microperthite in many specimens is kaolinized and consists of potash feldspar penetrated by slender stringers and irregular patches of oligoclase-andesine (Ab_{70}). Adjacent microperthite grains commonly show sutured boundaries. In addition to their presence in the microperthite, potash feldspar and oligoclase-andesine (Ab_{70}) also occur as subhedral and euhedral grains.

Quartz occurs interstitially, suggesting late crystallization.

Biotite is subhedral and euhedral, and some grains contain inclusions of apatite and zircon. Magnetite occurs both on the borders of and within the biotite grains. In a few specimens the borders of the biotite show alteration to

chlorite. A specimen of quartz syenite (Specimen MM 105, Table 6) from a point near the northern tip of the essexite body yielded biotite with the following optical properties: α = ca. 1.600, β = 1.656, γ = 1.657; optically negative; 2V = ca. 3°; X ∧ c = ca. 2°; dispersion r < v, weak; X < Y = Z, X = pale yellow Y and Z = orange brown. Judging from these optical properties, this biotite is composed mainly of the annite molecule. A specimen of somewhat finer-grained quartz syenite (Specimen MM 32, Table 6) from Willard Stream, north-northeast of the mountain summit, contains biotite with the following optical properties: β and γ = ca. 1.630; 2V = 13°30′; dispersion r < v, strong. This biotite probably has a lower ferrous iron-magnesia ratio than the preceding one.

Hornblende forms anhedral and subhedral grains intimately associated with biotite. According to Wolff (1929, p. 8), the hornblende from the quartz syenite of the Norton mine on the eastern edge of the stock has a specific gravity of 3.422 and optical properties as follows: α = 1.693, β = 1.711, γ = 1.713; 2V = 43.5°; negative; Z ∧ c = 20–21°; X = light yellow, Y = olive green, Z = deep grass green. The chemical analysis reported by Wolff (1929, p. 9) is shown in Table 2. As Wolff (1929, p. 15) points out, this mineral is actually hastingsite Billings, 1928).

Three specimens of quartz syenite from east of the essexite screen were examined. Both megascopically and microscopically these are the same as those from the main quartz syenite stock, except that they carry a small amount of sphene. Estimated modes are included in Table 6.

TABLE 6.—ESTIMATED MODES OF THE QUARTZ SYENITE IN VOLUME PER CENT

Specimen No.	MM1	MM2	MM4	MM7	MM9	MM12	MM17	MM24	MM30	MM32	MM50	MM59	MM60	MM105	MM107	MM110	MM112
Microperthite	75	60	70	70	70	80	70	75	75	70	73	65	75	70	70	59	66
Quartz	5	20	20	15	5	5	17	5	10	5	10	15	15	10	15	20	8
Biotite	5	5	3	3	10	5	5	5	2	5	2	3	2	5	3	10	10
Hornblende	5	5	2	2	5	5	1	5	3	5	5	2	2	5	3	1	1
Potash feldspar	5	5	3	5	5	3	3	5	5	5	5	10	4	5	4	5	5
Plagioclase	5	5	2	5	5	2	4	5	5	10	5	5	2	5	5	5	10
Totals	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100

Specimens number 1, 2 and 107 collected from east of the essexite screen; all others collected from main stock.

Accessory minerals usually amount to 1 per cent or less and have been omitted from the above modes.

Granite

The granite is typically tough, structureless, and brownish pink, with a tendency to glomeroporphyritic texture. It is very similar to the quartz syenite except that it is pinkish and has more quartz and fewer dark minerals. The granite is medium-grained with conspicuous feldspar laths, clots of quartz, and aggregates of dark minerals.

Microscopically, the granite is hypidiomorphic granular and consists essentially of microperthite, quartz, untwinned potash feldspar, plagioclase, hornblende, and biotite. The texture is equigranular and typically medium-grained, glomeroporphyritic. The grain size ranges from 0.5 mm to 3 mm and averages 1 mm. Accessory minerals are magnetite, apatite, allanite, and zircon. Biotite and hornblende are rarely altered to chlorite, but the perthite shows strong alteration to kaolinite.

Textural relations in the granite are the same as those in the quartz syenite, but there are a few striking mineralogical differences: (1) The microperthite is composed of potash feldspar and more sodic oligoclase (Ab_{78}). (2) Oligoclase (Ab_{78}) and potash feldspar also occur as separate grains. (3) The estimated modes (Table 7) show that the quartz content is generally higher and the dark minerals lower than those of the quartz syenite. (4) Pleochroism of the biotite is different from that of the biotite in the quartz syenite. It is strong reddish brown parallel to the cleavage and nearly colorless normal to the cleavage.

A study was made of a specimen from a granite dike (Specimen MM 89, Table 7) which

cuts the quartzite country rock. This granite is much finer-grained and averages about 0.4 mm in grain size. It contains less than 1 per cent biotite.

TABLE 7.—ESTIMATED MODES OF THE GRANITE IN VOLUME PER CENT

Specimen No.	MM 37	MM 67	MM 89
Microperthite	57	70	60
Quartz	27	20	25
Potash feldspar	3	2	5
Plagioclase	8	3	10
Biotite	5	3	tr.
Hornblende	tr.	2	tr.
Total	100	100	100

Specimen MM 89 is from a granite dike.
Accessory minerals usually amount to 1 per cent or less and have been omitted from the above modes.

STRUCTURE OF THE STOCK

General Statement

The Mount Monadnock stock is a composite body consisting of successive plutonic intrusions followed by a retinue of dikes.

Main Quartz Syenite Body

Medium-grained, pink, subporphyritic quartz syenite with biotite and hornblende makes up most of the main intrusive body. This is a tough, massive, extremely resistant rock, somewhat variable in texture and mineral content. Throughout, its quartz content ranges from nearly 0 to 25 per cent; the rock thus passes from true syenite through quartz syenite (nordmarkite) to granite. Many of these varieties appear to be small separate intrusions with sharp contacts. The granitic variety of quartz syenite differs from the other younger granite at Mount Monadnock which occurs in larger bodies and resembles closely the Conway granite of the White Mountain magma series.

The quartz syenite shows two other variations. Locally the grain size becomes less than 1 mm, and the rock can be classified as light-colored, fine-grained quartz syenite, but this change seems to be gradational. Exposures on the summit of the mountain near the fire tower are typical of this light-colored, fine-grained type.

Another variety of quartz syenite is the dark fine-grained type that is distributed widely but is not abundant. This is well displayed in an outcrop on the north bank of Willard Stream, N. 15° E. of the summit, where large irregular masses of it are clearly transected by apophyses of the more normal medium-grained pink quartz syenite. The dark fine-grained quartz syenite is clearly older here and probably forms large xenoliths. Similar relations may be seen along the margin of the stock on the crest of Todd Hill, due north of the summit of Mount Monadnock, as well as in many other places.

Because of great difficulty in locating boundaries, and because of sudden and repeated changes in texture and quartz content, all these syenitic rocks of the main body were mapped as one unit. The medium-grained quartz syenite is by far the most abundant rock in the main stock, and hence, for the sake of simplicity, throughout this paper the name quartz syenite will be used to embrace all the syenitic types described above.

Angular to subrounded masses of fine-grained, gray diotite are abundant in the quartz syenite at many places, and some may be seen on the trail leading to the summit of the mountain. These average 1–2 inches in diameter, but some are much larger. For example, in Willard Stream, N. 10° E. of the summit, a large block of pink syenite contains many angular inclusions of this diorite, some over a foot across. On the southeast slope of the mountain these inclusions appear at many points and range from 1 inch to 1 foot in diameter. The largest, somewhat less than 50 feet long, lies at the northern edge of the stock at the summit of Todd Hill.

Boundaries between the diorite masses and the enclosing rock are essentially sharp and usually no zone of transition is present, but in some places phenocrysts of feldspar, similar to those in the syenite, appear in the diorite, and stringers of quartz syenite cut it (Pl. 2, fig. 2). The diorite masses are believed to be remnants of an early diorite intrusion that were torn off and incorporated by the quartz syenite magma.

Angular xenoliths of schist and quartzite, from a few inches to several feet across, are abundant at many points around the margin of the quartz syenite stock, but most of these are

too small to be mapped. At the north end of the stock, however, on Todd Hill, three unusually large xenoliths of micaceous quartzite are mapped diagrammatically (Pl. 1). Other large xenoliths form the north end of the essexite screen.

Intrusion of Gray Porphyritic Syenite

A gray porphyritic syenite crops out sporadically from elevation 1740 feet to elevation 1880 feet in the stream which flows south from the mountain summit. The gray seynite is medium-grained with phenocrysts of white feldspar up to half an inch across. It is cut by a fine-grained pink leuco-syenite, and both rocks in turn are intruded by a thin pegmatite. The gray syenite contains abundant irregular inclusions, up to 8 inches in diameter, of black, fine-grained, biotite-rich rock which also carries large feldspar phenocrysts. Boundaries between inclusions and host rock are hazy, suggesting assimilation. The age sequence of the rock types, beginning with the oldest, is: (1) black fine-grained, biotite-rich rock, (2) gray porphyritic syenite, (3) pink leuco-syenite, and (4) pegmatite. The gray porphyritic syenite is the most voluminous, but, since its outline cannot be determined accurately, it had been mapped diagrammatically (Pl. 1).

Granitic Intrusions

Several small intrusions of medium-grained, pink biotite granite, resembling the Conway granite, perforate the earlier quartz syenite of the stock. The largest of these bounds the stock on the south as a long narrow body arranged parallel to the contact. At its west end this is clearly defined, but its east end is buried by glacial debris. Contacts between granite and quartz syenite were not seen here, but observations elsewhere indicate the granite is younger. Its apparent crescentic shape and its arrangement suggest that this intrusion is a small ring dike, the granitic magma having risen along the boundary of the quartz syenite stock.

Between elevations 1500 and 1900 feet on the northeast slope of the mountain, medium-grained, pink granite is exposed in several places. Outcrops are sparse and of poor quality, however, so that the distribution is not known precisely, and it is uncertain whether the granite forms a single stock or a group of smaller, perhaps dikelike bodies. Consequently, it is outlined on the map very diagrammatically.

One small intrusion of granite, barely large enough to map, lies at an elevation of 1300 feet due east of the summit, and another similar one at an elevation of 1400 feet on the steep slope S. 55° E. of the summit. Both of these are outlined diagrammatically. Several other irregular intrusions of granite, resembling the Conway granite, were noted, but these are too small to be mapped.

Screen

On the east slope of Mount Monadnock, between elevations 1500 and 2000 feet, is a screen composed of rock older than the quartz syenite. This slender arcuate body, with an average width of 500 feet, has been traced almost continuously throughout its length of 1¼ miles. The north end has been located rather accurately, but the south end is not exposed and may extend somewhat farther south. The exact location of the screen on the ridge N. 80° E. of the summit is uncertain since the topography here, between elevations 1700 and 2000 feet, is evidently inaccurate. The screen is composed principally of medium-grained, locally fine-grained essexite, but its extreme northern end consists of large, angular, randomly oriented blocks of fine-grained quartzite and schist, like the country rock surrounding the mountain, cut by stringers of quartz syenite. Heavy soil and vegetation prevented accurate delineation of the individual blocks in the field so they are shown diagrammatically on the geological map

---

PLATE 2.—MOUNT MONADNOCK, VERMONT

FIGURE 1. VIEW OF MOUNT MONADNOCK FROM A HILL 3 MILES TO THE EAST
Note flat essexite bench on south (left) flank of mountain.

FIGURE 2. BOULDER IN WILLARD STREAM ON NORTH SIDE OF MOUNT MONADNOCK
Stringers of medium-grained quartz syenite cut fine-grained diorite.

FIGURE 1

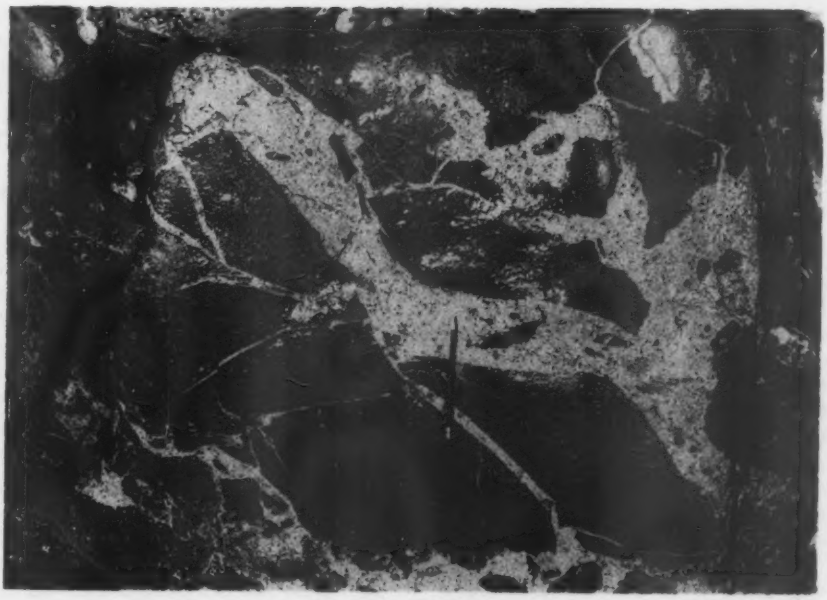
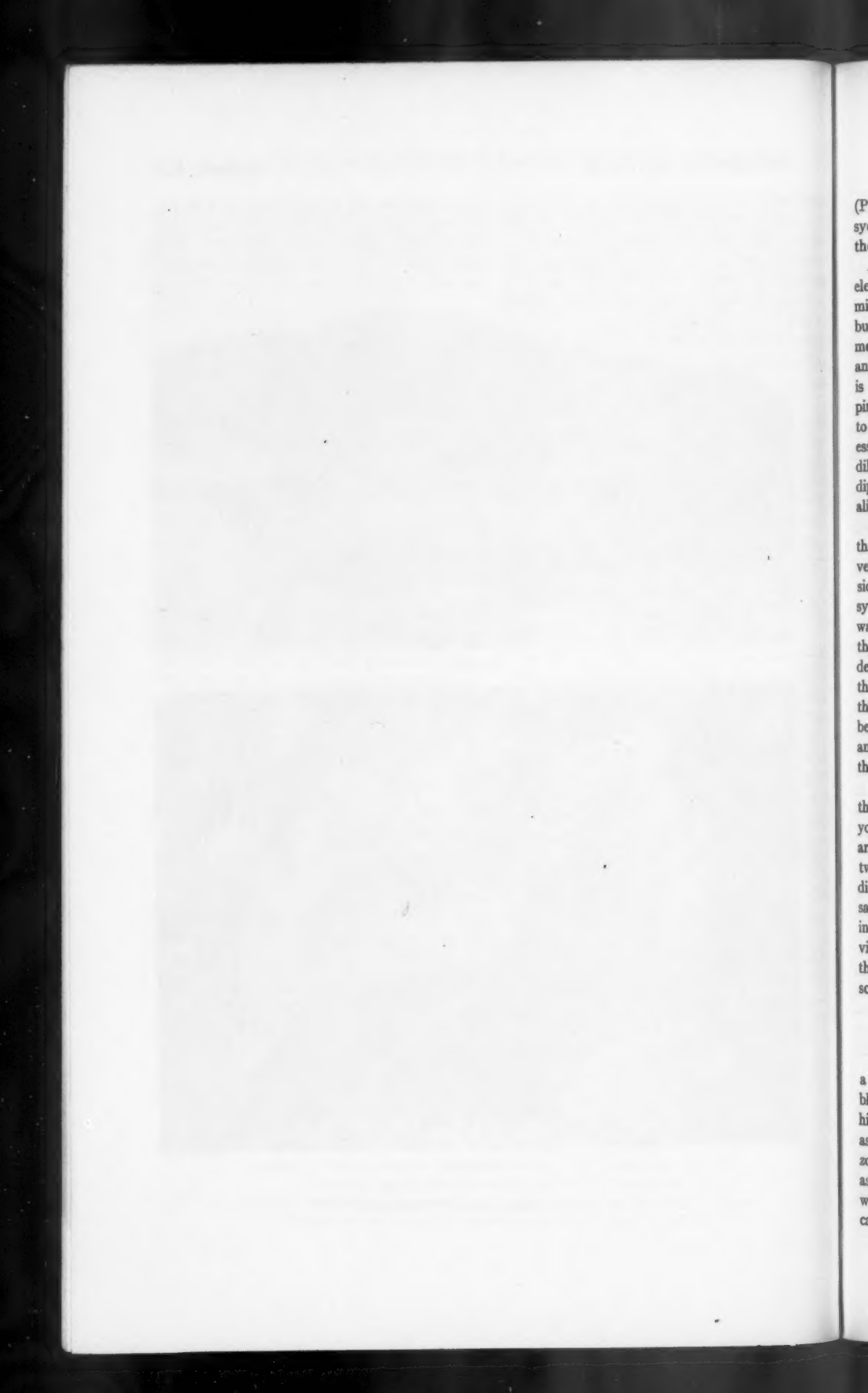


FIGURE 2

MOUNT MONADNOCK, VERMONT



(Pl. 1). The screen is terminated by quartz syenite immediately to the northeast and to the northwest.

A fine exposure of the screen may be seen at elevation 1670 feet, where the trail to the summit crosses the brook N. 70° E. of the summit, but the screen here is abnormally narrow. The medium-grained, dark-gray essexite, without any noticeable structural features except joints, is cut by irregular dikes of medium-grained, pink quartz syenite from a fraction of an inch to 2 feet wide. Abundant angular inclusions of essexite are engulfed in the syenite. A 3-foot dike of lamprophyre, striking N. 25° E. and dipping 85° NW., cuts essexite and syenite alike.

The strike of the screen essentially parallels that of the contact, although the north end diverges slightly toward the west. On its east side the screen is separated from the quartz syenite by a sharp, distinct boundary, but toward the west the essexite appears to grade into the rock of the transition zone, about to be described. Distribution of outcrops suggests that the east wall of the screen is steep, although this is not conclusive. The body is believed to be a giant slab of essexite with some quartzite and schist pried from the wall of the stock by the incoming quartz syenite magma.

This at once raises the question as to whether the quartz syenite east of the screen is older or younger than that to the west, or whether both are of the same age. In other words, are these two masses separate and distinct intrusions of different ages, or are they both parts of the same intrusion? No structural evidence exists indicating that the two masses are distinct individuals, and petrographic studies demonstrate that the syenitic rocks on either side of the screen are essentially similar.

Transition Zone

Immediately west of the essexite screen lies a zone of medium-grained, dark rock resembling the quartz syenite in texture but having a higher content of femags, in some cases as much as 15 per cent. Several traverses across this zone reveal that it has an outline approximately as shown on the geological map (Pl. 1) although within the zone quartz syenite appears sporadically. As already noted, the rock of the transition zone grades into the essexite of the screen. A similar relationship appears to exist between the dark transition rock and the quartz syenite to the west, although the gradation is complete within perhaps 100 feet or so. Petrographic studies suggest that the rock of the transition zone is a hybrid, formed when quartz syenitic magma became contaminated with essexite at the time of intrusion.

Dikes

The plutonic rocks of Mount Monadnock are cut by a retinue of late dikes covering a wide range of composition. These have been observed transecting the schist and quartzite surrounding the mountain and all the plutonics except the granite. It appears that the order of age of the dikes, from oldest to youngest, is syenite, granite, quartz bostonite, camptonite, aplite, and pegmatite. The dikes range in width from half an inch to about 10 feet, but in no case could the length be determined. Plate 1 shows only those hypabyssal intrusions the strike and dip of which could be measured accurately, but many others, with questionable attitudes, have been seen in the field. The dikes, on the whole, dip steeply and show a striking relationship to the boundary of the quartz syenite stock, each one striking either parallel or perpendicular to the boundary nearest its location.

Boundary of the Stock

Fortunately, the contact of the stock with the surrounding country rock (schist and quartzite) can be seen in many places, and the resulting observations furnish valuable data on the mechanics of stock emplacement. The boundary has been traced accurately around most of the intrusion, but along the southeast side it is buried by glacial drift and river alluvium. Although broadly elliptical, it is irregular in detail, and its position on the whole conforms rather strikingly to the topographic break in slope around the mountain.

Both the most accessible and revealing exposures are along that portion of the ridge of Todd Hill due north of the summit of Mount Monadnock (Pl. 1). The ridge east of the sharp bend in the road follows the contact approximately,

and geologic relations are complex. Both medium-grained quartz syenite and dark fine-grained quartz syenite are present; small dikes of the former penetrate the dark phase. Along the ridge to the southeast, outcrops of quartz syenite alternate with micaceous quartzite, and it is clear that the boundary is actually a zone marked by large xenoliths of metamorphosed country rock engulfed in quartz syenite and by dikes of quartz syenite cutting country rock. Angular xenoliths of fine-grained diorite are especially abundant here, and these range in diameter from 1 inch to over 1 foot. They are more numerous near the local summit of the ridge (elevation 1760 feet), and finally, about 100 feet south of the local summit, a long thin body of fine-grained diorite is exposed trending parallel to the ridge. This body, slightly exaggerated on Plate 1, is probably less than 50 feet long and is transected by narrow dikes of medium-grained pink quartz syenite. Undoubtedly, this rock has furnished the local inclusions of fine-grained diorite, and perhaps all the small xenoliths so abundant throughout the main quartz syenite stock were derived from this early rock type.

The western boundary of the stock evidently comes close to a point in Willard Stream N. 40° W. of the summit of Mount Monadnock, because good contact relations are exposed here. In a picturesque gorge, a 4-foot dike of medium-grained leucogranite (quartz-rich phase of the quartz syenite) penetrates schistose quartzite for a distance of 25 feet, and many small dikes of the same rock abound. Irregular xenoliths of quartzite, a foot or more across, are lodged in the large dike, where they have been contact metamorphosed to a dense hornfels.

To the south, at an elevation of about 2000 feet, in the stream valley N. 75° W. of the summit of the mountain, the boundary of the stock is also exposed. On the upstream (east) side are syenite and quartz syenite of the main part of the stock which give way, toward the contact, to quartzite cut by a large mass of quartz syenite. Farther downstream, *i.e.*, closer to the contact, a deep gorge exposes medium-grained, highly jointed, pink granite on its south side and foliated quartzite on the north. Granitic dikes penetrate far into the quartzite, and blocks of the latter are incorporated in the granite. Finally, downstream one passes into normal quartzite of the country rock. The granite has thus apparently insinuated itself along the contact zone which at this locality is probably about 200 feet wide.

This broad, irregular, zonelike character of the boundary of the stock is strikingly expressed along an old logging road that climbs the ridge up the southwest side of the mountain. At an elevation of 1750 feet, on a line extending S. 40° W. from the summit, outcrops of black fine-grained schist alternate with those of medium-grained to fine-grained quartz syenite as one proceeds across the boundary. Seemingly, the schist represents huge blocks detached and engulfed by the quartz syenite but not moved very far. The schistosity is highly crumpled and rather flat. Smaller xenoliths of schist also occur in this zone. Small intrusions of quartz syenite, ranging from mere stringers to dikes 6 inches across, cut through the schist. Most of these trend N. 40° E. and dip 55° NW.

On the ridge S. 15° W. of the summit of the mountain at an elevation of 1715 feet, a multitude of dikes, composed of medium-grained to fine-grained, white to pink granite, intrude schist. The granite is a quartz-rich phase of the syenite, one which is common, although not abundant, throughout the stock. The dikes are irregular, have variable strikes, and range in width from 1 inch to 1 foot, except that one is several feet across and has a questionable outline. These apophyseal intrusions, 750 feet southwest of the boundary of the stock as exposed at the surface, show that the magma penetrated the wall rock to a considerable distance.

The boundary of the stock is exposed on the south bank of Willard Stream N. 35° E. of the summit, and here, as in all other places, it is a zone rather than a line. Outcrops of medium-grained quartz syenite alternate with exposures of quartzite, and many dikes of the former cut the quartzite.

Six hundred feet east of the above outcrop and on the north side of the stream, an irregular 3-foot dike of quartz syenite cuts fine-grained black quartzite. It strikes N. 65° W. and dips 75° NE. The medium-grained quartz syenite is chilled to a finer-grained rock near the margins, and locally it passes into fine-grained granite.

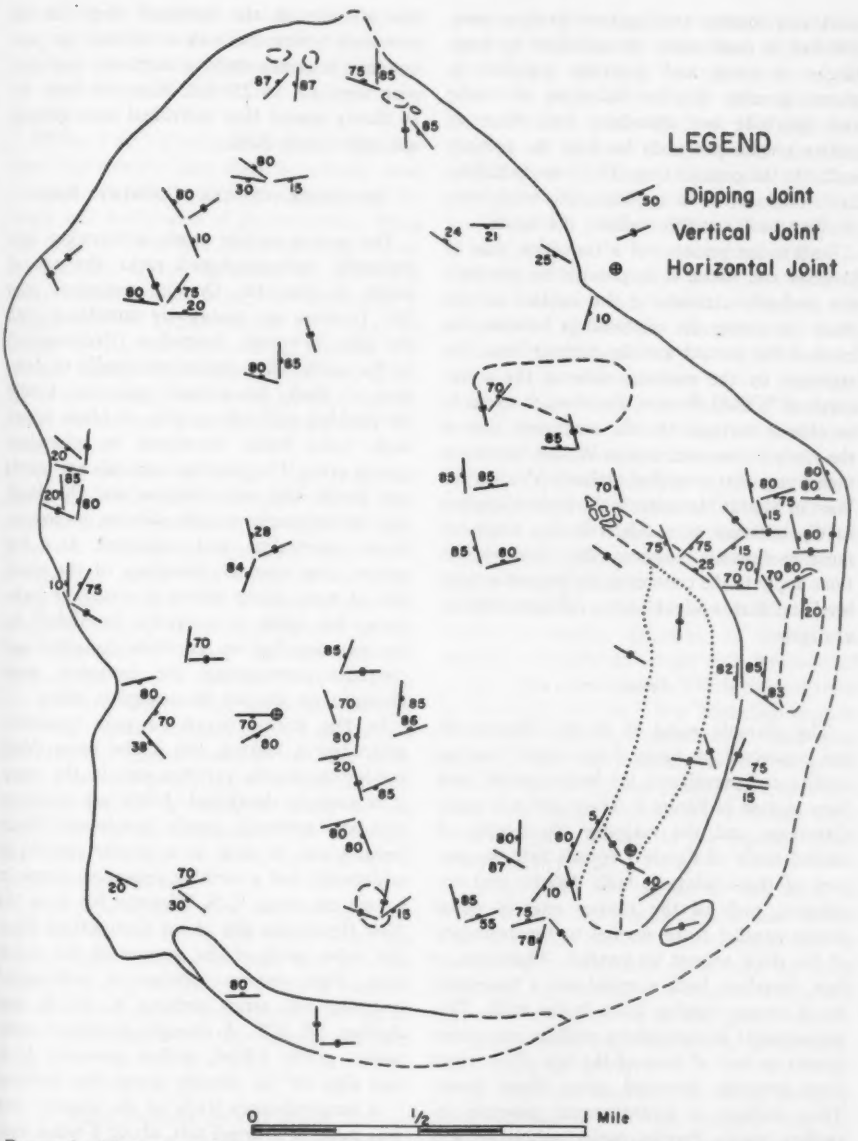


FIGURE 1.—MAP OF THE JOINTS IN THE QUARTZ SYENITE STOCK AT MOUNT MONADNOCK, VERMONT

An apophysis from this dike branches off to the south. It will be noted that this point is about 500 feet northeast of the stock boundary.

The eastern boundary of the stock was located approximately at an elevation of 1140 feet in the stream that flows eastward toward Colebrook, New Hampshire. Nowhere, however, was the boundary actually seen on the east or southeast sides of the mountain.

Thus, wherever seen the boundary between

stock and country rock appears to be a zone, 100 feet or more wide, characterized by large blocks of schist and quartzite engulfed in quartz syenite. Smaller inclusions of schist and quartzite are abundant, and dikes of quartz syenite penetrate far into the country rock. On the geologic map (Pl. 1) the boundary has been drawn so as to include this whole zone of dikes and xenoliths within the stock.

Despite the presence of a transition zone of breccias and dikes, it is possible to ascertain the probable attitude of the contact of the stock by noting the relationship between the trace of the contact and the contour lines. For example, on the northeast side of the stock, north of Willard Stream, the contact seems to be almost vertical. On the northwest side of the stock the contact crosses Willard Stream in such a way that a vertical attitude is suggested. West of Mount Monadnock the contact appears to dip outward at a relatively low angle. It must be realized, however, that these deductions apply to the contact at the present erosion level and that it might have a different attitude at depth.

Joints

The plutonic rocks of Mount Monadnock are transected by many joint sets of varying quality and prevalence, the best of which have been plotted in Figure 1. These strike in many directions, and the majority dip steeply. A careful study of Figure 1 reveals that the pattern of these joints is both definite and significant; each of the steeper ones is either nearly parallel to or normal to the boundary of the stock nearest its location. There are, in fact, therefore, both a radial and a tangential set of steeply dipping joints in the stock. This arrangement is particularly striking and corresponds to that of most of the late dikes which were probably intruded along these joints. Thin coatings of hydrothermal minerals including quartz, fluorite, pyrite, sphalerite, and molybdenite are present on some joint faces.

Gently-sloping joints, dipping from 5 to 25°, were found at many points in the stock, but these are less common than the steep joints. Distribution of the measured ones is shown in Figure 1.

Well-developed sheeting occurs along the two streams on the southeast slope of the mountain where the rock is divided by pronounced fractures striking northeast and dipping from 10° to 25° SE. Many of these are so closely spaced that individual slabs of rock are only 1 inch thick.

Structure of the Country Rock

The quartz syenite stock is intrusive into regionally metamorphosed rocks the age of which is uncertain. On the geological map (Pl. 1) these are tentatively correlated with the Gile Mountain formation (Ordovocian?) to the south. They consist principally of dark-gray to black, fine-grained quartzite, locally interbedded with silvery-gray to black schist, both rocks being transected by abundant quartz veins. The principal minerals are quartz and biotite with some feldspar and chloritoid, and the accessories include chlorite, muscovite, zircon, tourmaline, and magnetite. At a few points, very near the boundary of the stock, and at some places where it occurs as inclusions, the schist is somewhat hornfelsed by the intrusion, but on the whole the schist and quartzite surrounding the intrusion show virtually no contact metamorphic effect.

In the thicker layers of pure quartzite, schistosity is lacking, but in the more thinly bedded micaceous varieties and in the schist it is strongly developed. Joints are relatively rare and generally poorly developed. Where bedding can be seen, it is usually parallel to schistosity, but a striking exception occurs in a road cut along U. S. Highway No. 3 on the New Hampshire side of the Connecticut River 5½ miles north of the summit of the mountain. This outcrop consists of well-bedded quartzite and schist striking N. 55° E. and dipping 30° NW. A strongly developed schistosity, gently folded, strikes generally N.-S. and dips 60° E. directly across the bedding.

A reconnaissance study of the country rock was made in a broad belt, about 5 miles wide, around the mountain, and far beyond the limits of Plate 1. It was found that the regional strike of the schistosity in the outer part of this belt averages N. 15° E., and the dip about 30° SE. Exceptions do occur, as along the road from Canaan to Averill, 5½ miles north-northeast of the mountain, where the strike ranges

from N.-S. to N. 40° W. and the dip is toward the northeast. A more thorough study would be required to work out the detailed structure of the country rock, but the regional trend is obvious.

Within 1 or 1½ miles of the intrusion, however, the country rock departs radically from the regional strike and dip, especially on the north and south sides of the mountain. Along the ridge of Todd Hill just west of the stock, for example, where the rocks are somewhat more schistose than usual, the schistosity is highly deformed and abnormally flat. The strike swings through all points of the compass due to the shallow dip. Stratification and schistosity are parallel, and in places both are intensely contorted. Fold axes plunge from 15° to 25° NE.

A traverse up the long ridge projecting south from Mount Monadnock shows relations similar to those on Todd Hill. The rock here is a schistose quartzite with considerable interbedded schist. The strike of the schistosity ranges widely, and in many places it is nearly horizontal. At an elevation of 1500 feet the schistosity is highly crinkled, and the folds plunge 30° in a direction S. 20° E. Close by, however, similar folds are nearly vertical.

In Mill Brook, to the west, the quartzite is schistose and highly crinkled, and the schistosity dips very gently. More or less the same relations prevail in Blodgett Brook to the southwest. On the west side of the mountain the strike is uniformly N.-S., but the dip is gentle toward the east.

Clearly the country rock close to the stock, especially on the north and south sides, has been disturbed much more than that at a distance of from 2 to 5 miles, and it is believed that this local intense disturbance may be, in some way, related to the intrusion of the stock.

Emplacement of the Rock

The mode of emplacement of the intrusions of the White Mountain magma series, to which series the rocks of Mount Monadnock belong, is an intriguing problem. Probably most of the ring dikes, like those in the Tertiary igneous province of Great Britain, and many of the stocks originated by simple sinking of large cylindrical blocks of country rock into a reservoir of magma, the outlines of the blocks later being modified by piecemeal stoping. However, some of the ring dikes appear to have resulted primarily from the piecemeal stoping of magma along arcuate fractures (Williams and Billings, 1938, p. 1041; Billings, 1945, p. 44–55; 1947, p. 286; Chapman, 1942, p. 1556–1558; 1948, p. 1093). There is little, if any, evidence in the areas studied thus far that the country rock was either forcefully pushed aside by the magma or that it was completely replaced by "granitizing" mechanisms. As it happens, Mount Monadnock yields considerable evidence on the structural features of an intrusion of the White Mountain magma series, and the bearing of this evidence on the mode of emplacement will now be considered. Nine significant points are involved:

(1) The plutonic rocks of Mount Monadnock are typical representatives of the White Mountain magma series, quite lacking in foliation and lineation.

(2) In ground plan the Mount Monadnock pluton is roughly elliptical, its major axis making a considerable angle with the regional strike of the country rock. This is characteristic of the stocks of the White Mountain magma series. By itself, the cross-cutting relationship suggests that the magma was not squeezed forcibly into its present position, for, if it had been, the intrusion more likely would have been elongated parallel to the regional trend of the country rock.

(3) In detail the outline of the stock is irregular, and the boundary is a zone rather than a definite surface. The country rock is penetrated by abundant dikes from the main intrusion, and xenoliths of diorite, schist, and quartzite are numerous in the quartz syenite. Obviously, piecemeal stoping was active especially along the contact.

(4) Wherever observed, all actual contacts between igneous rock and country rock are knife sharp rather than gradational, and the country rock, except as inclusions, never contains crystals of feldspar or quartz that might have been derived from the igneous rock. There is no good evidence whatever that the rocks of the stock represent metamorphic rock that was "granitized" by magma or by emanations of plutonic origin.

(5) The screen of essexite, which seems to have been wedged from the east wall of the stock, is a feature characteristic of cauldron subsidence; its almost exact counterpart, composed of volcanic rocks, occurs in the syenite stock of Ascutney Mountain, Vermont (Chapman and Chapman, 1940). There is a strong indication that the magma separated this arcuate slab from the wall rock by inserting itself along a broadly circular fracture formed by ring-fracture stoping.

(6) The elongate intrusion of granite along the southern edge of the quartz syenite stock is shaped like a ring dike. Even though its eastern end is not exposed, the body appears to be arcuate and wedged in between the main stock and the country rock. This evidence does not imply any specific method of intrusion, however, since the granite may have reached its present position in one of several ways such as by forcing itself along the contact, by stoping its way upward, or by flowing upward passively as the walls of a fracture became separated. The disposition of the three smaller granite bodies to the east and north of the essexite screen suggests that all may have been emplaced along the same circular fracture which was concentric with the eastern boundary of the stock.

(7) The plutonic rocks possess two well-defined, persistent sets of joints, one radial and one tangential. Both sets are vertical or steeply dipping. Coatings of hydrothermal minerals suggest that these joints formed immediately upon consolidation of the magma.

(8) Late dikes, covering a wide compositional range, cut the plutonics and the surrounding metamorphic rocks. Those of which the attitude could be determined show both radial and tangential patterns like the joints.

(9) The strikes and dips of the country rock immediately north and south of the intrusion differ radically from those at greater distance. The well-developed schistosity, so persistently uniform in a zone 2 to 5 miles away, has been crumpled and tilted close to the mountain until it is nearly horizontal. Near the mountain, fold axes range widely both in direction and degree of plunge even within a small radius. The question arises as to whether this complex local disturbance is merely the result of a late phase of regional deformation, or whether it was produced by the quartz syenite magma as it forced its way into place.

It is clear that the rocks composing the Mount Monadnock stock, like those forming other plutons of the White Mountain magma series, were crystallized from intruded magma and do not represent granitized country rock. Cauldron subsidence, accompanied by the stoping of large, arcuate slabs, seems certainly to have been the major process of intrusion as evinced by the screen of essexite and by the ring dike of granite. The elliptical outline of the stock with its major axis forming a considerable angle with the regional schistosity lends additional support to this interpretation.

It is thought that the mechanism of intrusion probably operated somewhat as shown in Figure 2. A roughly domical reservoir of differentiating magma lay at shallow depth beneath the surface. For some reason the roof above this reservoir became unstable, and it settled, thereby developing a set of steep, cylindrical fractures and a contemporaneous set of radial ones. These did not necessarily reach the surface, however, even though the roof itself was probably relatively thin. As the fractures formed, arcuate slabs of roof rock began to founder, and the magma rose to take their place. Perhaps a large rock cylinder foundered first, giving the stock its generally circular ground plan, and, later, large arcuate slabs and small irregular blocks were stoped off the walls, thereby widening the body horizontally. The essexite screen presumably represents one of these slabs composed of an early differentiate of the magma.

Settling of the roof evidently continued over a considerable period of time, even after the upper part of the magma body had crystallized. This newly consolidated portion, subject to stresses of similar nature but of lesser intensity, developed steep arcuate faults and steep radial and tangential joints. Granitic differentiates rose along some of the arcuate faults where they consolidated as thin crescentic ring dikes. Hydrothermal solutions and late hypabyssal dikes used the radial and tangential joints, as avenues of easy access during the final stage of consolidation.

The explanation outlined above differs somewhat from those proposed by Anderson (1936),

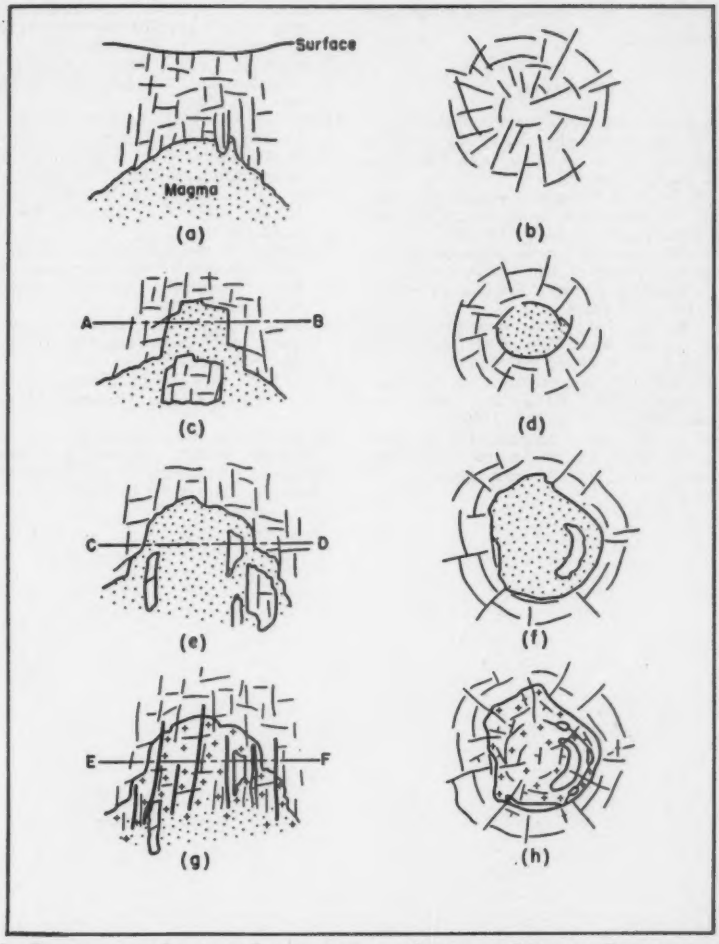


FIGURE 2.—DIAGRAMMATIC SKETCHES ILLUSTRATING THE SUGGESTED MODE OF EMPLACEMENT OF THE ALKALINE STOCK AT MOUNT MONADNOCK, VERMONT

(a) Roof sags into magma reservoir developing faults and fractures. (b) Map of (a) showing radial and concentric fracture patterns. (c) A cylindrical block settles into the magma. (d) Horizontal section along line A-B in diagram (c). (e) Arcuate slabs are stoped from the reservoir walls. (f) Horizontal section along line C-D in diagram (e). (g) Roof and crystallized upper portion of magma (shown by crosses) subside, steep arcuate faults and radial and tangential joints form, and granite ring dikes and late hypabyssal dikes are intruded. (h) Map of present surface, drawn along line E-F in diagram (g), showing quartz syenite stock with screen, joints, granite ring dikes, and late hypabyssal dikes.

Richey (1932), and Billings (1945, p. 44–55; 1947, p. 284–288) for cauldron subsidences in general.

Presumably the passive intrusive mechanism suggested here for Mount Monadnock resulted when the pressure of the magma became greatly reduced. Such appears to be the normal situation in areas of cauldron subsidence. However, north and south of the stock the quartzite and schist have been strongly twisted and their schistosity thrown out of the regional strike. It is suggested that, during the very early stages of intrusion, positive magma pressure was sufficiently vigorous, for a brief period

of time, to deform the immediately surrounding metamorphic rocks.

References

Anderson, E. M., 1936, The dynamics of the formation of cone sheets, ring dikes, and cauldron subsidences: Royal Soc. Edinburgh Yearbook, v. 56, pt. 2, p. 128–157.

Billings, M. P., 1928, Chemistry, optics, and genesis of the hastingsite group of amphiboles: Am. Mineralogist, v. 13, p. 287–296.

—— 1945, Mechanics of igneous intrusion in New Hampshire: Am. Jour. Sci., v. 243-A, p. 40–68.

—— 1947, Structural geology: New York, Prentice-Hall, Inc., 473 pages.

Chapman, R. W., 1942, Ring structures of the Pliny region, New Hampshire: Geol. Soc. America Bull., v. 53, p. 1533–1568.

—— 1948, Petrology and structure of the Percy quadrangle, New Hampshire: Geol. Soc. America Bull., v. 59, p. 1059–1100.

Chapman, R. W. and Chapman, C. A., 1940, Cauldron subsidence at Ascutney Mountain, Vermont: Geol. Soc. America Bull., v. 51, p. 191–212.

Chapman, R. W. and Williams, C. R., 1935, Evolution of the White Mountain magma series: Am. Mineralogist, v. 20, p. 502–530.

Eggleston, J. W., 1918, Eruptive rocks at Cuttingsville, Vermont: Am. Jour. Sci., v. 45, p. 377–410.

Greenwood, R., 1951, Younger intrusive rocks of Plateau Province, Nigeria, compared with the alkalic rocks of New England: Geol. Soc. America Bull., v. 62, p. 1151–1178.

Richey, J. E., 1932, Tertiary ring structures in Britain: Geol. Soc. Glasgow, Trans., v. 19, p. 42–140.

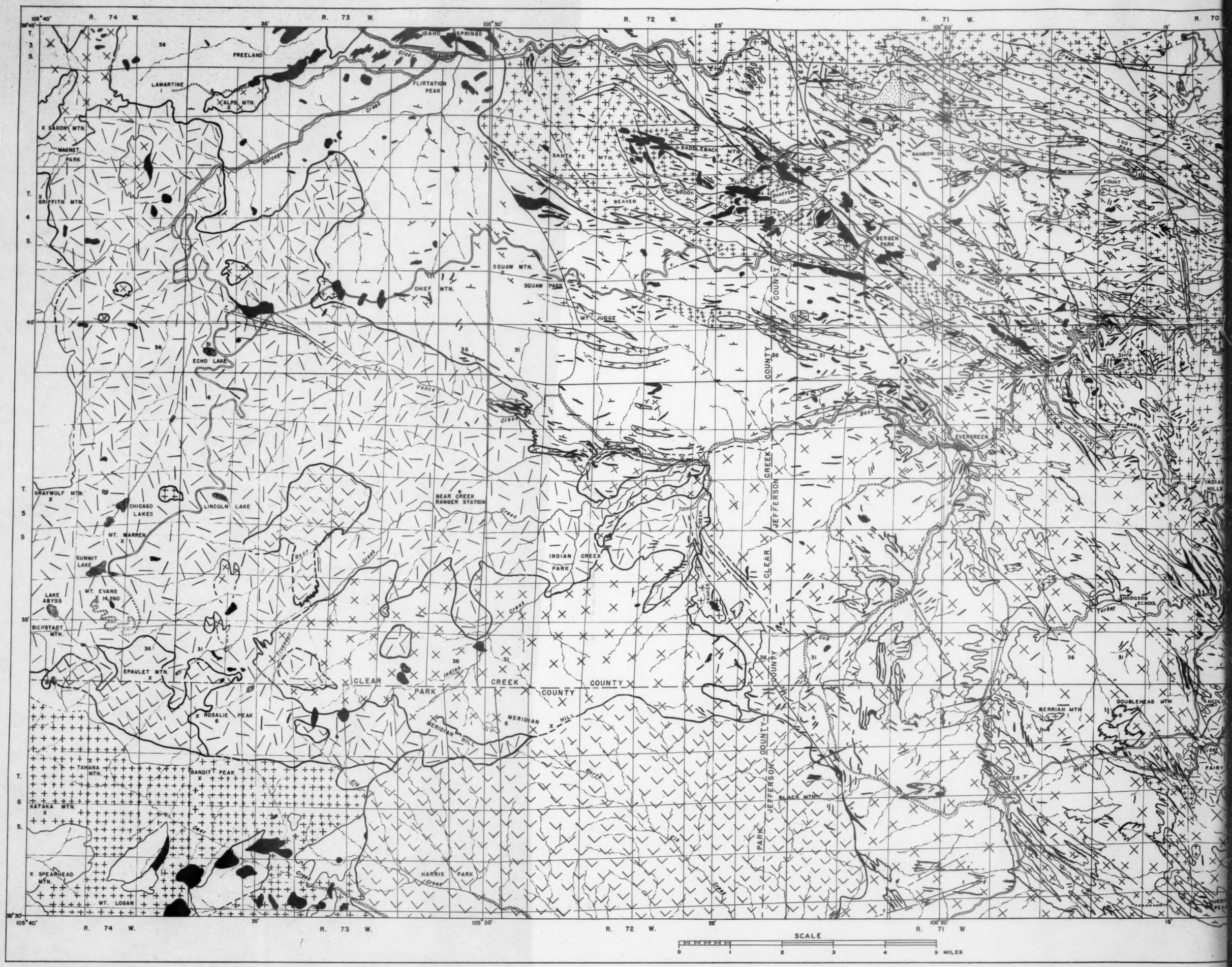
Williams, C. R. and Billings, M. P., 1938, Petrology and structure of the Franconia quadrangle, New Hampshire: Geol. Soc. America Bull., v. 49, p. 1011–1044.

Winchell, A. N., 1951, Elements of optical mineralogy, pt. II: New York, John Wiley and Sons, Inc., 551 pages.

Wolff, J. E., 1929, Mount Monadnock, Vermont—A syenite hill: Jour. Geology, v. 37, p. 1–16.

United States Geological Survey, South 157 Howard Street, Spokane 4, Washington

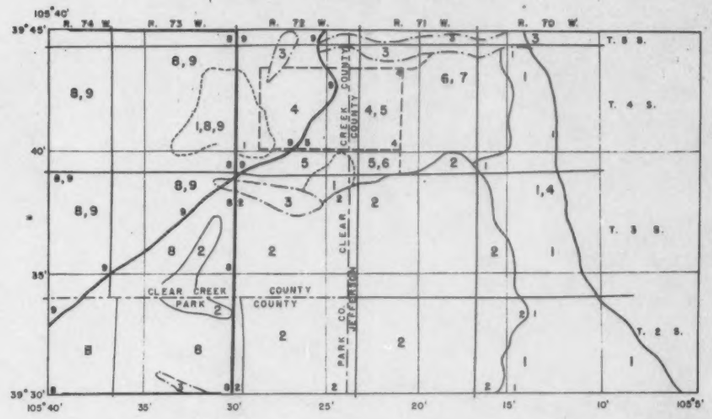
Manuscript Received by the Secretary of the Society, December 16, 1952



GRANITE PEGMATITES AND RELATED IGNEOUS ROCKS OF THE DENVER MOUNTAIN PARKS AREA, COLORADO

INDEX MAP

SHOWING SOURCES OF GEOLOGIC DATA



FIELD GEOLOGY BY

1. MARGARET FULLER BOOS 1936 - 1950
2. M. F. BOOS & E. ABERDEEN 1937 - 1939
3. M. F. BOOS & C. M. BOOS 1950
4. HELEN D. VARNES 1939 - 1941
5. H. D. VARNES, W. KRAXBERGER & M. F. BOOS 1939
6. JACK AMES, SAM DALLISON & M. F. BOOS 1940
7. ROBERT MORTON, EDWARD LEMON, L. SALETEL & M. F. BOOS 1941
8. BALL, SPURR & GARREY, U.S.G.S. PROF. PAPER 63 PLATE II 1908
9. LOVERING & GODDARD, GEOLOGIC MAP OF THE FRONT RANGE MINERAL BELT, COLORADO, 1939 P.P. 214, 1951

EXPLANATION

PLEISTOCENE

MORAINE

PRECAMBRIAN

SILVER PLUME AND MOUNT OLYMPUS GRANITES

PIKES PEAK GRANITE

DIORITE AND METADIORITE

BOULDER CREEK GRANITE AND GRANITE GNEISS

MOUNT MORRISON ORTHOGNEISS MIGMATITE AND INJECTION GNEISS

IDAHO SPRINGS AND SWANDYKE FORMATIONS UNDIFFERENTIATED

CONTACT

CONTACTS OCCUPIED, IN PART, BY PEGMATITE

FAULTS AND BRECCIA REEFS

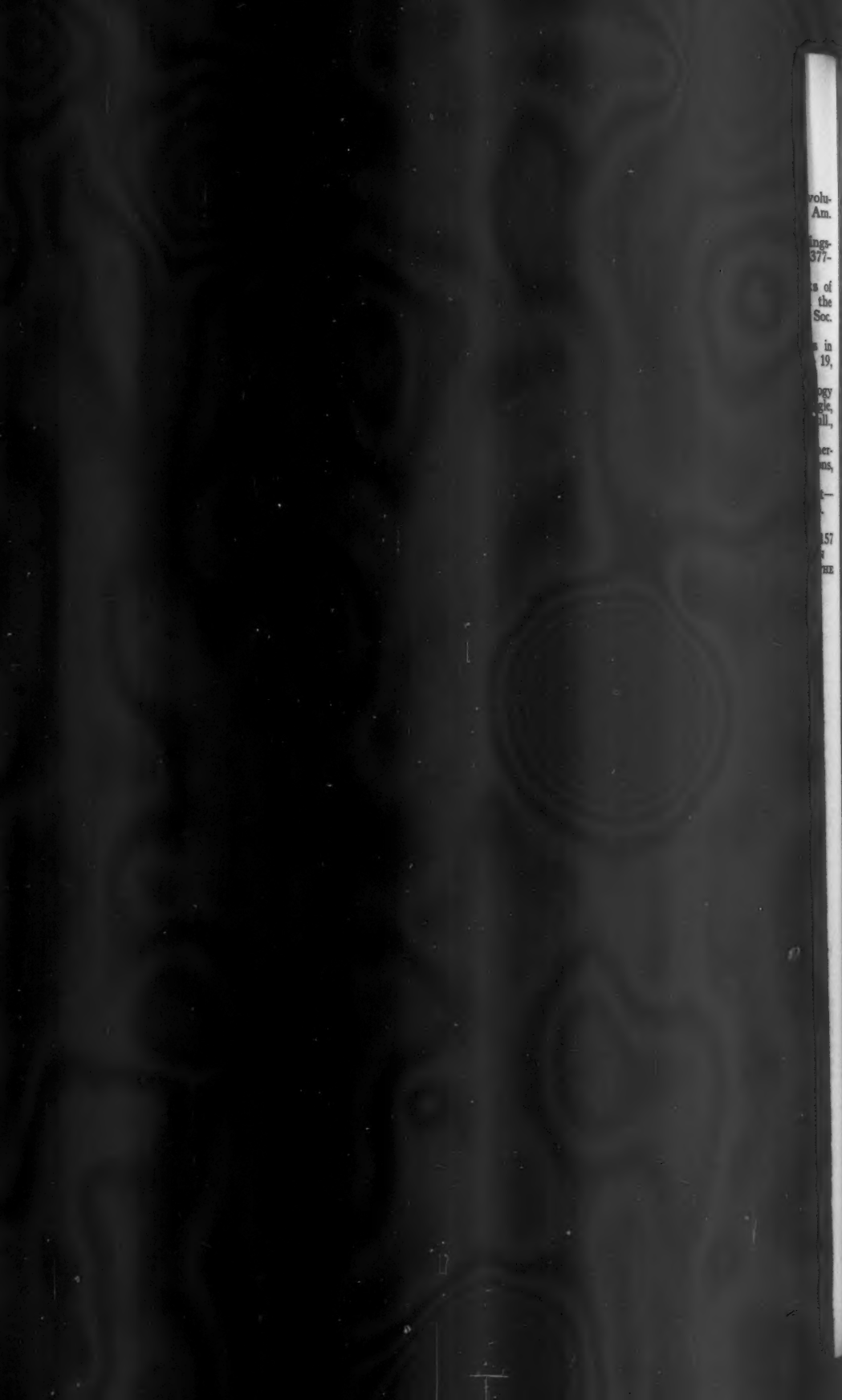
FAULTS OCCUPIED, IN PART, BY PEGMATITE

PEGMATITE AND STRONGLY PEGMATITIC GRANITE

SCALE

0 1 2 3 4 5 MILES

GRANITE PEGMATITES AND RELATED IGNEOUS ROCKS OF THE DENVER MOUNTAIN PARKS AREA, COLORADO



BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 65, PP. 115–142, 8 FIGS., 5 PLS. FEBRUARY 1954

GENESIS OF PRECAMBRIAN GRANITIC PEGMATITES IN THE DENVER MOUNTAIN PARKS AREA, COLORADO

BY MARGARET FULLER BOOS

ABSTRACT

Pegmatites genetically related to the Mount Morrison gneiss and migmatite, the Boulder Creek granite and granite gneiss, a quartz diorite, the Pikes Peak granite, and the Silver Plume and Mount Olympus granites invaded the Denver Mountain Parks area, particularly the Idaho Springs and Swandyke formations in the order named, probably in late Precambrian times.

The development of the five granitic rocks was accompanied by thousands of pegmatites in the form of dikes, sills and irregular bodies. Field studies indicate that the pegmatites of each granite surge are distinctive. The characteristics that distinguish one granite from another are reflected, to a limited extent, in the pegmatites genetically related to each granite cycle.

Relatively few pegmatites are complex. Some appear to have been subjected to more than one pegmatization. Some dikes, initiated as simple primary pegmatites in the earlier granite cycles, were replaced and enriched during one or more of the succeeding periods. Fault and shear zones served as thoroughfares for repeated mineralization of pegmatite bodies that partially filled them.

CONTENTS

TEXT

	Page
Introduction	116
Location of the area	116
Field work	116
Acknowledgments	117
Regional geologic setting	117
General geology	117
Metamorphic rocks	117
Granitic rocks	118
General statement	118
Mount Morrison gneiss and migmatite	118
Boulder Creek granite gneiss	119
Quartz diorite	119
Pikes Peak granite	119
Silver Plume granite	119
Mount Olympus gray granite	120
Interrelationships of granitic and metamorphic rocks	120
Pegmatites	120
General	120
Distribution of pegmatites	122
Structural relations of pegmatites, aplites and host rocks	124
Problems of pegmatite genesis	125
Primary pegmatite identities	126
Characteristic features of primary simple pegmatites and the granites of each cycle	126
Mount Morrison cycle	126
Boulder Creek cycle	127
Pikes Peak cycle	128
Silver Plume-Mount Olympus cycle	129
Quartz diorite	130
Summary	130
Complex pegmatites	131
Geologic associations of the complex pegmatites	131
Genesis of certain complex pegmatites	137
Genesis	139
Answers to questions	139
References cited	140

ILLUSTRATIONS

Figure	Page
1. Location of the area	116
2. Pegmatite-rich areas and location of pegmatites of Table 1	118
3. Map of Grover pegmatite and mine	132
4. Maps of Bigger pegmatite and mine	133
5. Map of pegmatites on East Shaffer Hill	133
6. Map of Coors pegmatites and mines	134
7. Map of Sunrise Peak pegmatite and mine	135
8. Map of Wasson pegmatite and mine	136

Plate	Facing page
1. Granite pegmatites and related igneous rocks of the Denver Mountain Parks area, Colorado	115

	Following page
2. Mount Morrison gneiss, migmatite, injection gneiss and pegmatites	122
3. Boulder Creek granite gneiss and pegmatite, quartz diorite, Pikes Peak granite, Silver Plume granite and Mount Olympus granite	
4. Pegmatites exposed in the rugged terrain of Denver Mountain Parks	
5. Special features of the pegmatites	

TABLE

1. Pegmatites that have been mined or prospected in the Denver Mountain Parks area	124

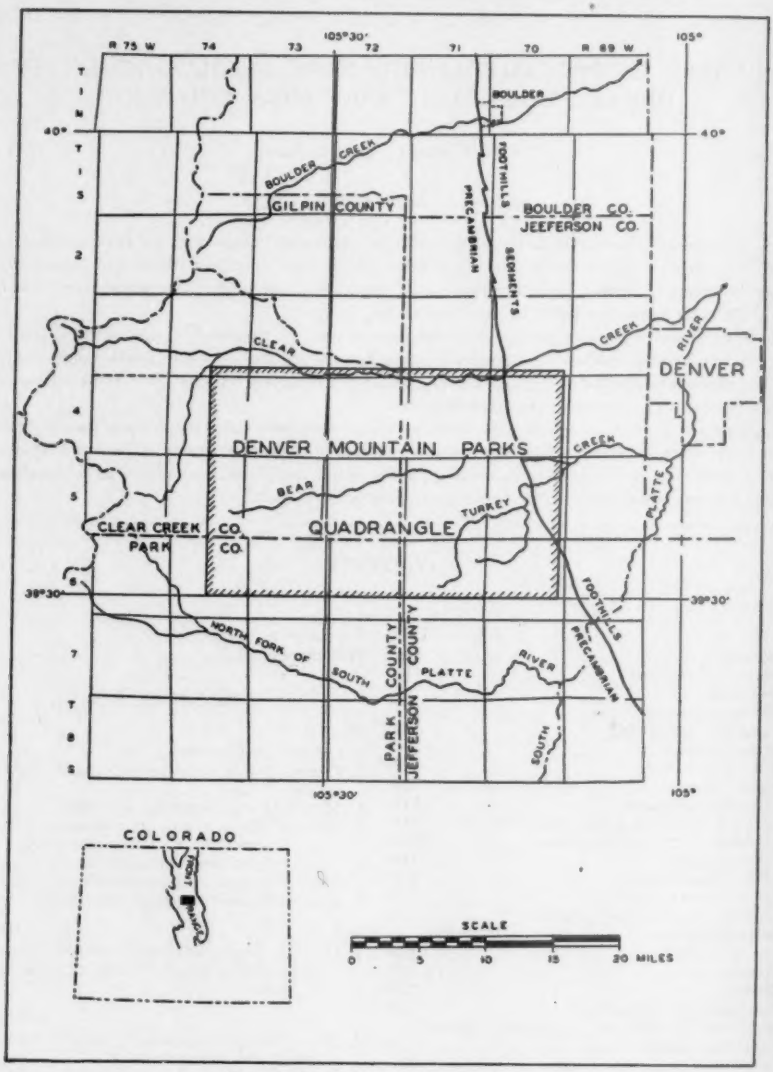


FIGURE 1.—LOCATION OF THE AREA

INTRODUCTION

Location of the Area

The Precambrian granitic pegmatites of the Denver Mountain Parks area outcrop on the eastern flank of the Front Range west of Denver, Colorado (Fig. 1). The pegmatites considered here outcrop chiefly in the western two-thirds of Jefferson County, in the southeastern part of Clear Creek County, and in the northeastern part of Park County. The topography is shown on the Denver Mountain Parks, Morrison, and Littleton quadrangles.

Field Work

Field examination of the pegmatites began in 1935. The investigations of the Indian Creek plutons and associated rocks included the peg-

matites (Boos and Aberdeen, 1940). Study of the Precambrian extended through 1942. The work was recessed in 1943 and 1944 but was resumed in 1945 along with investigations of the structural geology of the area (Boos, 1947, p. 1179; 1949, p. 1874).

The interpretation of pegmatite genesis is based on careful field studies of approximately 3,000 bodies of pegmatite and aplite. Three-dimensional data were obtained by examination of numerous prospects and the year-to-year mapping of mined pegmatites.

The observations resulted in a considerable body of information which is presented in the pages that follow. Further mineralogical studies, now in progress, are expected to substantiate or modify the conclusions reached.

Acknowledgments

Students at the University of Denver assisted in the field from 1935 to 1943. Much of the work was checked in the field by the author. Space does not permit individual identification of and acknowledgment to all who worked on the problems for brief periods.

Geologists who assisted materially with the field work include Dr. Esther Aberdeen, C. Maynard Boos, Helen D. Varnes, John Ames and Sam Dallison. The author only, however, is responsible for the interpretations given and the conclusions reached.

Field and laboratory studies have been aided by two grants from the Penrose Bequest of The Geological Society of America. A grant in 1940 was devoted entirely to studies of the Precambrian pegmatites and related rocks. One phase of the work in 1936–1937 pertained to the pegmatites and was aided by a grant from the Society. Sincere thanks are extended for the generous assistance.

Regional Geologic Setting

General Geology

The formations exposed on the eastern flank of the Front Range in the Denver Mountain Parks area consist chiefly of Precambrian granitic rocks, paragneisses and paraschists, and associated meta-igneous formations. Repeated disturbances have greatly folded and intensely altered the rocks.

Field observations indicate that the granitic rocks occupy part of a large comagmatic region that extends beyond the boundaries of the Front Range. In the Denver Mountain Parks area there are extensive exposures of at least five generations of granitic rocks and their pegmatites and aplites. The granitic rocks compose about 75 percent of the crystalline rocks exposed (Fig. 2).

The pegmatites occur within the granites and in the metamorphic rocks that enclose granites. The distribution of the granitic rocks has exerted a marked control on the concentration and character of the pegmatites (Pl. 1). The crystalline mosaic is complex. The pegmatites also interpenetrate intricately.

Metamorphic Rocks

The metamorphic rocks include the Idaho Springs series that was derived chiefly from the alteration of thick sediments, and the Swandyke gneiss of igneous origin. The metamorphic rocks are the oldest formations of the area and are widely exposed in a broad belt that extends east and southeast from Idaho Springs 12–15 miles and disappears eastwards under the sedimentary formations of the foothills bordering the Front Range on the east (Pl. 1).

The type locality of the Idaho Springs series is near Idaho Springs, about 30 miles west of Denver (Spurr, Garrey and Ball, 1908, p. 37). In many places, the metamorphic rocks of the Idaho Springs series interfinger so intimately with small bodies of granite and granite gneiss that the details cannot be shown on the maps that accompany this paper. Areas of igneous rocks, other than pegmatites, having less than a quarter of a square mile of outcrop are not shown on Plate 1 except where closely related to large numbers of pegmatites.

The Idaho Springs consists of several formations whose areal and stratigraphic limits have not yet been defined in the Front Range. Intense isoclinal folding, minute crumpling, and complex injection by several generations of igneous rocks have so disturbed and altered the formations that the thicknesses and stratigraphic boundaries of individual units are difficult to correlate.

Descriptions of the Idaho Springs beds given by Ball (Spurr, Garrey and Ball, 1908, p. 38) apply, in general, to the gneisses, schists, and

related rocks in the area. The lower portion of the Idaho Springs here consists of quartz-biotite schist and quartz-biotite-sillimanite schist interbedded with quartz gneiss and lime-silicate and carbonate rocks (Boos, 1951, p. 1533). The upper part consists of stratified hornblende gneiss interlayered with quartz gneiss, mica schist, and lime-silicate and carbonate rocks.

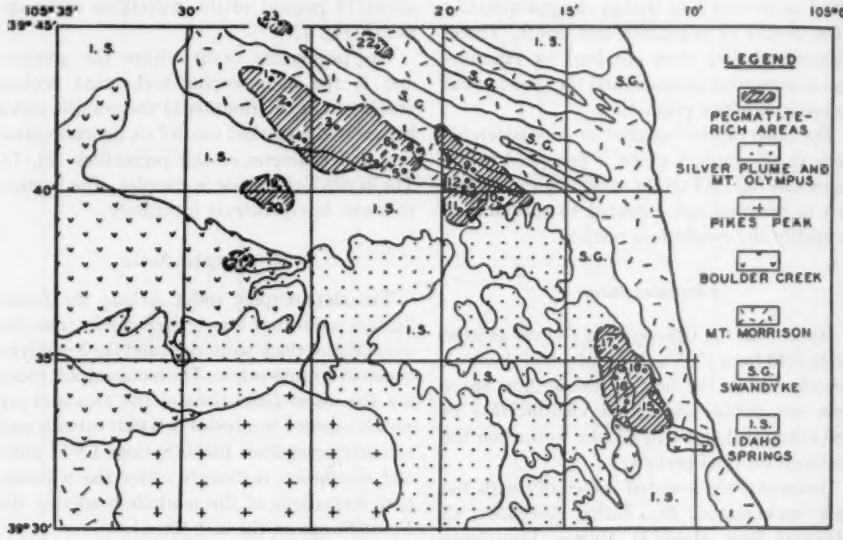


FIGURE 2.—PEGMATITE-RICH AREAS AND LOCATION OF PEGMATITES OF TABLE 1

The Swandyke hornblende gneiss (Lovering, 1935, p. 1) and related pegmatites outcrop east and northeast of Idaho Springs (Map of Front Range mineral belt, Lovering and Goddard, 1939, 1951). The hornblende gneiss overlying the Idaho Springs series in the northeastern part of the Denver Mountain Parks area and the numerous basic dikes that cut the Idaho Springs beds are probably the equivalent of the Swandyke at the type locality in the Montezuma area.

The meta-igneous hornblende gneiss of probable lava flow origin in the Swandyke cannot everywhere be distinguished with certainty in the field from the metasedimentary hornblende gneiss and associated carbonate rocks in the upper part of the Idaho Springs proper. Locally, the Swandyke is unconformable on the Idaho Springs, and the gneissosity and general appearance of the two hornblende gneisses differ considerably at their junction. Dikes of hornblende gneiss cross the Idaho Springs beds.

Granitic Rocks

General statement.—The granitic rocks of the area consist, from older to younger, of Precambrian quartz monzonite gneiss (orthogneiss of Lovering and Goddard, 1951) and related pegmatites; Boulder Creek granite, granite gneiss, and related pegmatites; small bodies of quartz diorite; Pikes Peak granite and related pegmatites; Silver Plume granite, and Mount Olympus granite and related pegmatites. The granitic rocks are close counterparts of the granitic rocks north, south and west of the area mapped in the Front Range (Spurr, Garrey and Ball, 1908, p. 40–45; Boos and Boos, 1934, p. 304; Lovering and Goddard, 1938, p. 11–16; 1951, geologic map; Bastin and Hill, 1917, p. 33–37; Boos and Aberdeen, 1940, p. 700).

Mount Morrison gneiss and migmatite.—The earliest granitic rock consists of finely gneissic quartz monzonite, migmatite, and injection gneiss with associated pegmatites and aplites. The chief rock is a gneiss that occurs as imbricated sheetlike masses and stocks. The boundaries and gneissosity are essentially parallel to the foliation of the enclosing metasedimentaries.

Large bodies are well exposed on Santa Fe and Saddleback Mountains southeast of Idaho Springs (Pl. 1). Stocks of the migmatitic rock compose most of Mount Morrison and Mount Falcon in the eastern part of the area near the foothills. Small irregular masses of it outcrop in the Deer Creek drainage area. The stocks are interconnected by sheets and sills that are interlayered with the Idaho Springs and Swandyke beds in the northeastern part of the area.

Ball (Spurr, Garrey and Ball, 1908, p. 47) first described the rock as a quartz monzonite gneiss (Ademellite gneiss) and gneissoid granite; Lovering and Goddard (1939 map, 1951, p. 23) designated it as quartz monzonite gneiss. In neither case, however, did this earliest granitic rock receive a formation name. So widespread a formation deserves a designation. It is here referred to as the Mount Morrison formation. The type locality is on Mount Morrison where a large body of it is well exposed.

The Mount Morrison formation is distinctly gneissoid due to the parallel elongation of clusters of biotite flakes alternating with aggregates of quartz and feldspar (Pl. 2, figs. 1, 2). The rock is essentially a migmatite in the eastern part of the area. In general, injection gneiss borders the Mount Morrison migmatite and grades into it. One phase of the gneiss is rich in augen centered by biotite-magnetite spots.

The Mount Morrison gneiss is medium-grained, and on fresh exposure is gray to almost white or tan. The granulation of the quartz and feldspar is striking. All phases have a pronounced light-and-dark banding with light minerals composing 80–90 percent of the gneiss. Foliation follows the banding and has persistent trend for long distances. The gneissosity is dragged, swirled, and contorted against non-granitic rocks in many places. Inclusions of quartz-mica schist and hornblende gneiss are numerous in some outcrops.

The formation contains a wealth of pegmatitic and aplitic material that is so much like the gneiss itself that it is difficult to distinguish the dikes from the main body of the Mount Morrison. Apophyses of pegmatite extend into the adjacent formations for long distances, becoming aplitic distally.

Boulder Creek granite gneiss.—Small bodies of this granite gneiss outcrop in the eastern part of the area, but the main mass is centered in the Mount Evans batholith (Boos and Aberdeen, 1940, Pl. 1). The Boulder Creek is not a strong dike- or sill-maker. The irregular plutons of dark, fine-to-medium-grained gneissic diorite and metadiorite that occur in the Turkey Creek area are probably of Boulder Creek age (Pl. 1).

Typical Boulder Creek granite and granite gneiss is gray to blue-gray, coarse-textured, subporphyritic, unevenly granular, and massive to strongly gneissic (Pl. 3, figs 1, 2). In places the rock is dark and strongly foliated. Pegmatites genetically related to the Boulder Creek granite gneiss are not numerous. They occur in the country rock near the gneiss and in the granite gneiss and granite in places on Mount Evans (Pl. 5, fig. 2), on Fairy Hill, and between Sweitzer Gulch and the foothills.

Quartz diorite.—Small bodies of quartz diorite have a limited outcrop in Deer and Turkey Creek canyons. The rock is medium-grained, dark gray, unevenly granular, and not gneissic. Prominent hornblende crystals lie in a matrix of gray feldspar. Weathered surfaces, pitted where the feldspars have broken down, are unlike those of any other gray granitic rock in the area (Pl. 3, fig. 3).

Pikes Peak granite.—This granite, exposed in the south-central part of the map area, is part of the Rosalie lobe of the Pikes Peak batholith (Boos and Aberdeen, 1940, p. 699). Small bodies of it intrude the Mount Morrison gneiss and the Boulder Creek granite gneiss and are intruded by the Silver Plume granite (Spurr, Garrey and Ball, 1908, Pl. 11). Pods, lenses or dikes of Pikes Peak granite are very rare (Pl. 1).

The Pikes Peak granite is brick red or gray. A groundmass of gray or red orthoclase feldspar and microcline, with interstitial milky white quartz, is spotted with clusters of biotite and hornblende (Pl. 3, fig. 4). Pikes Peak granite is not gneissic. Pegmatites related to it occur sparingly in the granite, along the margin of the Rosalie lobe, and, scantily, in the adjacent gneisses and schists.

Silver Plume granite.—This granite composes the Indian Creek plutons (Boos and Aberdeen, 1940, p. 679) and intrudes the Pikes Peak granite and older formations. The Indian Creek plutons outcrop in a broad arc marginal to the northern end of the Pikes Peak batholith (Boos and Aberdeen, 1940, Pl. 1). Dikes and sills of

Silver Plume granite are numerous, locally, in the Idaho Springs formations.

The Silver Plume granite correlates physically, structurally, mineralogically and temporally with the Silver Plume of the batholith of that name, and with the chief granite of the Longs Peak-St. Vrain batholith. It composes most of the youngest Precambrian batholiths of the Front Range (Boos and Boos, 1933, p. 303–322; Boos, M. F., 1936, p. 67).

The rock is light gray to flesh-colored and decidedly fresh in appearance. It has conspicuous twinned, tabular microcline crystals that have preferred parallel orientation, producing a trachytoid texture (Pl. 3, fig. 5). The Silver Plume granite shows well-defined flow structures (Boos and Aberdeen, 1940, p. 707–712).

Mount Olympus gray granite.—The Mount Olympus granite was first recognized as a satellite of the Longs Peak (Silver Plume) granite in the Longs Peak-St. Vrain batholith (Boos and Boos, 1934, p. 322). It also forms a satellitic border to the Silver Plume granite of the Indian Creek Plutons. In both areas the Mount Olympus granite is a gray to white, "pepper and salt," even-grained rock that occurs in dikes and sheets that traverse the Silver Plume granite and adjacent rocks (Pl. 3, fig. 6).

Both Silver Plume and Mount Olympus granites are accompanied by many pegmatites and aplites that penetrate the older granitic rocks and the Idaho Springs and Swandyke formation. Wide zones of injection gneiss, rich in pegmatites, border the Indian Creek plutons in many places.

The distinctive physical, mineralogical and structural features of the five successive granitic formations permit identification and correlation of widely-separated outcrops. The abundant pegmatite and aplite bodies that accompanied each granite also have distinctive intrinsic features.

The granitic rocks penetrate the Idaho Springs and Swandyke formations and each other. The resulting crystalline mosaic, in some areas thickly strewn with pegmatites and aplites, is exposed as shown in Plate 1.

Interrelationships of Granitic and Metamorphic Rocks

The Idaho Springs formations were somewhat disturbed and altered before the granitic rocks invaded the area. The Mount Morrison gneiss transformed the Idaho Springs and Swandyke formations extensively and intensively. The considerable isoclinal folding that occurred during the development of the Mount Morrison is evident in the tight folds of the northeastern part of the area (Pl. 1). Some folding was coincidental with the emplacement of the Boulder Creek granitic rock in the Mount Evans batholith and the small stocks of the Turkey Creek area. Uplift and warping recurred when the great Pikes Peak batholith developed at the south, and again when the Indian Creek plutons of Silver Plume and Mount Olympus granite invaded the area around the north end of the Pikes Peak batholith.

Granitization and migmatization of existing rock by the Mount Morrison probably spread simultaneously from the Santa Fe Mountain-Saddleback and Mount Morrison-Falcon centers. Sheets originating in them interconnect in the northeastern part of the Denver Mountain Parks area, where they are interlayered with Idaho Springs and Swandyke beds (Pl. 1).

The gneissic margins of the Boulder Creek granite grade into the paragneisses and paraschists around them. The Pikes Peak granite contact is generally sharp against host rocks. The magma seems to have stoped its way into the area from the south, engulfing the earlier granites and the metamorphics.

The Indian Creek plutons of Silver Plume and Mount Olympus granite are generally accommodated to the foliation of schists and gneisses enclosing them. Both granites invaded older granite bodies.

PEGMATITES

General

Simple primary pegmatites (Landes, 1933, p. 33–56) constitute all but a very small fraction of the granitic dike and sill rocks genetically related to the late Precambrian granites. Only a very few of the thousands of pegmatities in the area are in any way complex (Turner and Verhoogen, 1951, p. 328).

The recurrent igneous activity, however, created a very complex record of pegmatite development. Pegmatite, pegmatite-aplite, and pegmatite-aplite-granite bodies, in which the

proportions of pegmatite, aplite, and granite are extremely variable, occur in profusion.

The earliest igneous rock developed in the area was the Swandyke gneiss, a massive basic rock which had its own pegmatites. No attempt is made in this paper to describe or delineate the characteristics of the Swandyke pegmatites or aplites.

Four of the five granites and granitic rocks already described briefly were accompanied by a great array of pegmatites, some of which preceded, some of which accompanied, and many of which followed the solidification of the larger granite bodies. The wealth of pegmatites and aplites within the granites, marginal to them, and in the schists and gneisses encompassing the granitic rocks, indicates that the acid pegmatites are closely related to the granites. Each granite cycle began with pegmatite and aplite, was accompanied by pegmatite of the giant granite type, and closed with the development of numerous bodies of pegmatite or pegmatite-aplite that developed chiefly from magma and rest magma. The profusion of pegmatites and their great variety of shape, size and structure stems from the long and recurrent pegmatitization of the area by the successive granitic invasions of late Precambrian times.

Pegmatites and aplites occur together everywhere, in many places as parts of the same granitic body. No distinction is made between them in this paper, so far as relationships to parent granite are concerned. In general, aplites are less numerous and smaller than pegmatites. Many sheets of general aplite aspect are streaked and spotted with pegmatitic material. Blebs of pegmatite that are welded to the aplite around them may be the result of the crystallization of the fluidal residua of magmatic material. In some places the patches of pegmatite appear to have resulted from the replacement of aplite. Elsewhere, the patches of pegmatite appear to be coarsely recrystallized aplite. Aplite may be the "mother rock" of considerable pegmatite (Uspensky, 1943).

Pegmatites within granite and indigenous to it are simple, giant granite structures containing many of the same accessory minerals that occur in the granite. The production of primary simple granitic pegmatites, wherever favorable physical and chemical conditions prevailed, occurred with the production of the finer-grained granite.

In the earlier granite cycles, especially the Mount Morrison, the first pegmatites formed along with the migmatization of the country rock and the injection gneiss that borders the numerous granite plutons of the area.

The pegmatite-producing agent in the later part of the early cycles and all through the late cycles (such as the Silver Plume-Mount Olympus) was essentially magmatic. Simple fractional crystallization of granite fluids enclosed in a relatively tight host rock resulted in differentiation, zone by zone, within a pegmatite body (Hanley *et al.*, 1950, p. 8–9; Cameron *et al.*, 1949, p. 16 *et seq.*). The development of the larger granitic bodies such as batholiths, stocks, or irregular plutons by magmatic action, metasomatism, or a combination of the two, resulted in an end product that was essentially fluid and crystallized. This produced the well-zoned pegmatites characteristic of the end of each granite cycle, with the exception, perhaps of the Mount Morrison cycle.

Because the Silver Plume-Mount Olympus cycle was from beginning to end essentially magmatic (Boos and Aberdeen, 1940, p. 729) its early, middle and late pegmatites have characteristic zonation and accessory minerals containing volatile compounds, lack replacement minerals and units, and show close association with the parent granite bodies.

The primary simple granitic pegmatites belonging to any granite cycle are light-colored, coarse-textured dikes and sills consisting of quartz, one or more feldspars, accessory minerals and some mica. No two of them are identical in form, shape, mineralogical organization and orientation, number and location of inclusions, and wall relationships to host rocks. More of them are massive than are definitely zoned or layered. Some vary little in mineral composition from one end of the exposed pegmatite body to the other. Others range, within a few feet on strike or dip, from monomineralic- to polymineralic-zoned and back to massive. There are hundreds of dikes and sills a few feet long and a few inches thick but relatively few pegmatites that are hundreds of feet long and tens of feet thick.

The very few pegmatites that are complex contain rare mineral units, and most of them also carry subcommercial bodies of mica, beryl, or feldspar (Table 1).

Distribution of Pegmatites

The Denver Mountain Parks terrain is rugged, its relief amounting to thousands of feet. Along numerous cliffs, canyons and steep mountain slopes that expose nearly vertical sections of Precambrian rocks, the three-dimensional aspects of pegmatites and asso-

---

PLATE 2.—MOUNT MORRISON GNEISS, MIGMATITE, INJECTION GNEISS, AND PEGMATITES

FIGURE 1.—MOUNT MORRISON GNEISS, MOUNT MORRISON

This texture is characteristic of the gneiss in the Rampart Range, between Mount Vernon canyon and Turkey Creek canyon.

FIGURE 2.—MOUNT MORRISON INJECTION GNEISS, FALCON MOUNTAIN

Note pegmatitic material developed by ultrametamorphism at the left, the ptygmatically folded, probably palingenetic pegmatite near the middle of the view, and the "flamen" type of metasomatic pegmatite along the axis of a small fold and blended into the gneiss, at the right.

FIGURE 3.—MOUNT MORRISON GNEISS AND PEGMATITE EAST OF THE "CASTLE" ON FALCON MOUNTAIN

The coarse-textured, somewhat gneissic pegmatite lies *lit-par-lit* with the foliation of the gneiss.

FIGURE 4.—MOUNT MORRISON GNEISS AND MIGMATITE

With a cross-cutting pegmatite, Olinger Hill, N. ½ sec. 22, T. 5 S., R. 70 W. The pegmatite, of late Mount Morrison age, is quartz rich and crudely zoned.

FIGURE 5.—LENS OF QUARTZ-RICH PEGMATITE WHICH BLENDS INTO ITS MOUNT MORRISON HOST ROCK IN PLACES

The pegmatite, of late Mount Morrison age, may have been developed metasomatically following the emplacement and solidification of the gneiss.

FIGURE 6.—"FLAMEN" TYPE PEGMATITE WITHIN INJECTION GNEISS OF MOUNT MORRISON AGE

The irregular branching knot of ill-defined pegmatite has developed in the axis of a minor fold. Pegmatites of this type are not zoned, usually carry relicts of host rock, and dwindle down into isolated porphyroblasts of feldspar.

PLATE 3.—BOULDER CREEK GRANITE GNEISS AND PEGMATITE, QUARTZ DIORITE, PIKES PEAK GRANITE, SILVER PLUME GRANITE, AND MOUNT OLYMPUS GRANITE

FIGURE 1.—BOULDER CREEK GRANITE GNEISS ON FAIRY HILL, SEC. 9, T. 6 S., R. 70 W.

Note the late Boulder Creek pegmatite, flecked with biotite, that crosses the gneissosity of the host on the hanging wall and is blended into the darker gneissic Boulder Creek below on the foot wall. Small late Boulder Creek pegmatite at extreme right.

FIGURE 2.—ZONED PEGMATITE OF POST-BOULDER CREEK GENESIS

The dike fills a fracture crossing the left-to-right foliation of rock that is probably a very dark phase of the Mount Morrison. Note the quartz-rich core at (Q), the quartz feldspar wall zone at (W) and the narrow border zones at (B). The gneissosity of the host rock is reflected in the pegmatite at (G).

FIGURE 3.—QUARTZ DIORITE EXPOSED IN TURKEY CREEK CANYON, SEC. 15, T. 5 S., R. 70 W.

The black spots are shadows in the pits where feldspars have weathered out. Light streaks are aplitic.

FIGURE 4.—PIKES PEAK GRANITE, BLACK MOUNTAIN, SEC. 13, T. 6 S., R. 71 W.

Note the haphazard arrangement of the clots of dark minerals.

FIGURE 5.—SILVER PLUME GRANITE

Note the parallel alignment of the light-colored, twinned microclines. Interstitial material is quartz, muscovite, biotite, and a very little plagioclase.

FIGURE 6.—MOUNT OLYMPUS GRANITE EXPOSED SOUTHEAST OF DOUBLEHEAD MOUNTAIN IN SEC. 8, T. 6 S., R. 71 W.

The outcrop has the characteristic "pepper-and-salt" appearance of typical Mount Olympus granite but shows considerable more biotite than is usually present in the Indian Creek plutons.

FIGURE 7.—GRANITE RELATIONSHIPS, OLD TURKEY CREEK ROAD

A Mount Olympus granite dike (B) occupies the contact of Boulder Creek granite at (A) with Boulder Creek granite gneiss at (C) that is injected with aplitic material.

FIGURE 1

FIGURE 2

FIGURE 3

FIGURE 4

FIGURE 5

FIGURE 6

MOUNT MORRISON GNEISS, MIGMATITE, INJECTION GNEISS, AND PEGMATITES

FIGURE 1

FIGURE 2

FIGURE 3

FIGURE 4

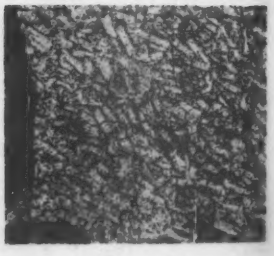


FIGURE 5

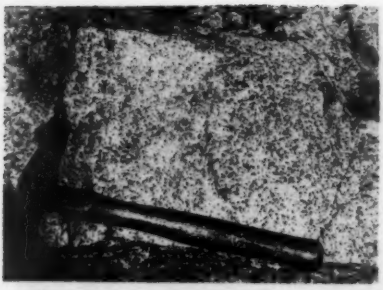


FIGURE 6

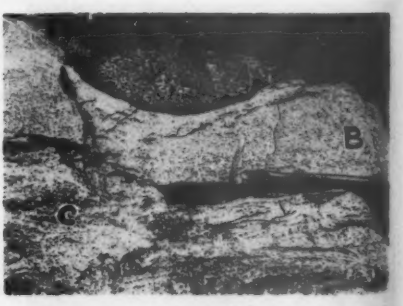


FIGURE 7

BOULDER CREEK GRANITE GNEISS AND PEGMATITE, QUARTZ DIORITE, PIKES PEAK GRANITE, SILVER PLUME GRANITE, AND MOUNT OLYMPUS GRANITE

FIGURE 3

FIGURE 1

FIGURE 2

PEGMATITES EXPOSED IN THE RUGGED TERRAIN OF DENVER MOUNTAIN PARKS

FIGURE 1

FIGURE 2

FIGURE 3

FIGURE 4

FIGURE 5

SPECIAL FEATURES OF THE PEGMATITES

ciated rocks are excellently displayed (Pl. 4, figs. 1, 2, 3).

Pegmatites and aplites occur in swarms, in groups of parallel dikes and sills, as cross-cutting bodies, in stockwerks, and *lit-par-lit* with the original bedding and present foliation of schists and gneisses. Many areas of granitic rocks are relatively free from pegmatite structures (Pl. 1). Very few outcrops of Idaho Springs and Swandyke beds lack pegmatites. An exception is the biotite-sillimanite member, on Bergen, Hicks and Judge mountains, which is relatively free from other granitic bodies. In general, there are few pegmatites in the mica schists except where granites have invaded them.

Pegmatites are more numerous adjacent to granitic rock than they are within the granite itself. Plutons, stocks, and batholiths of granite are fringed by pegmatites that inject the surrounding gneiss and schist. Swarms of dikes and sills occur in the environs of granitic masses.

In nongranitic host rocks, the Mount Morrison pegmatites are more numerous and widely distributed than are the pegmatites of later granitic cycles. Mount Morrison pegmatites were the voluminous and extensive invaders of the Idaho Springs and Swandyke formations. Many of them conform to the foliation (bedding) of the folded Idaho Springs beds. Northwest strikes and steep northeast dips are common. The pegmatites fill cross joints in the Swandyke gneiss. Dikes that trend northeast are rare. The pegmatite sills were plicated and crumpled along with the folds of the Idaho Spring beds in which they lie.

Pegmatites genetically related to the Mount

PLATE 4.—PEGMATITES EXPOSED IN THE RUGGED TERRAIN OF DENVER MOUNTAIN PARKS

FIGURE 1.—SOUTHEASTERN SLOPE OF MOUNT MORRISON

The country rock is Mount Morrison migmatite and injection gneiss crossed at (A) and (B) by late Mount Morrison pegmatites. Gullies are marked by lines of shrubby sage and follow the northwest-southeast foliation of the gneiss. Each juniper tree is about 15 feet high.

FIGURE 2.—PEGMATITES IN THE IDAHO SPRINGS SOUTH OF CLEAR CREEK

A swarm of pegmatites crosses the Idaho Springs beds and the Swandyke gneiss at (P) and extends across Clear Creek (C). The tributary valley at (T) is entrenched along a fault.

FIGURE 3.—VERTICAL SECTION OF PEGMATITE EXPOSED ON BEAR CREEK HIGHWAY

The dike is nearly vertical, about 25 feet high, and branches. The pegmatite crosses the foliation of the quartz-mica gneiss that is riddled with small vein-type ultrametamorphic pegmatites that follow the gneissosity.

PLATE 5.—SPECIAL FEATURES OF THE PEGMATITES

FIGURE 1.—PEGMATITES (P) ENCLOSING ROUNDED BODIES OF SILVER PLUME GRANITE (G) AND ENCLOSED BY SILVER PLUME GRANITE

FIGURE 2.—BOULDER CREEK PEGMATITE CROSSING BOULDER CREEK GRANITE GNEISS, MOUNT EVANS HIGHWAY ABOVE ECHO LAKE

The dike consists of coarse-textured potash feldspar and plagioclase with some quartz. There is practically no zonation, and the pegmatite is flecked throughout with small books of fresh black biotite.

FIGURE 3.—POD OF PEGMATITE IN THIN-BEDDED MARBLE MEMBER OF THE COLD SPRINGS GULCH MARBLE SERIES IN THE IDAHO SPRINGS FORMATION, BEAR CREEK

The pegmatite body is faintly gneissic, is not zoned, and has had little reaction with the adjacent, highly calcareous rocks.

FIGURE 4.—OPEN CUT IN BIGGER PEGMATITE, 1953

The view is southeast along the strike of the pegmatite. A "horse" of quartz crosses the dike from right to left (Q)-(Q), and two large prisms of beryl, 11 feet long by 24 inches thick and 9 feet long and 25 inches thick, were removed in 1939—*see* (B) in view.

FIGURE 5.—SEGMENTS OF THE BERYL CRYSTALS TAKEN FROM LOCATION (B) IN FIGURE 4 OF THIS PLATE

The pieces, scored by muscovite wedges, are from the upper end of the 11-foot-long prism. Almost all the beryl was removed before 1941, and the area that was formerly the site of the beryl crystals, and the bismuthinite adjacent to them, is now buried under quartz fragments.

Morrison formation appear in swarms south and southeast of Beaver Brook, especially adjacent to hornblende gneiss inclusions in the Mount Morrison. Pegmatites follow the axis of the east-west trending fold on the north slope of Santa Fe Mountain. They are abundant where the sills of Mount Morrison gneiss interfinger with the lime-silicate and carbonate beds in the Bear Creek area. Injection gneiss bordering Mount Falcon and Mount Morrison is rich in pegmatitic material. The migmatite and injection gneiss exposed on the north and east slopes of Mount Morrison is more than half pegmatite. Wherever the Mount Morrison outcrops, pegmatites abound.

Aplites are numerous among the granitic rocks of Mount Morrison age. Many of them are extensions of granitic sheets that branch from large bodies of migmatite and injection gneiss.

Pegmatites that stem from the Boulder Creek granite and gneiss occur singly, and only rarely in swarms. Many fill finite breaks in the Boulder Creek granitic rock on Mount Evans (Pl. 5, fig. 2). Others extend into the Idaho Springs beds nearby and cross the gneiss and schist at high angles. Dikes within granite usually consist of aplite streaked with pegmatite. There are only a few pegmatite-rich areas adjacent to outcrops of Boulder Creek granite gneiss.

The crystalline complex of the Denver Mountain Parks area was fairly rigid by Boulder Creek times. The primary pegmatites of Boulder Creek genesis are large, well-defined, tabular, cross-cutting bodies accommodated to the walls of the fractures they occupy in the biotite-sillimanite schist and quartz-mica gneiss. Pockets and streaks of pegmatite lie in the granite gneiss on Mount Evans.

The Fairy Hill-Medlen School stock of Boulder Creek gray gneiss (Fig. 2) is fringed with many of its pegmatites. Pegmatites of more than one granitic cycle are exposed along the old Turkey Creek highway, where their relative ages are plainly visible. The succession of granitic rocks is also well displayed (Pl. 3, fig. 7).

The Pikes Peak granite was not a great producer of pegmatite. The dikes have a characteristic red-and-white color and are easily recognized where they lie in the spotted Pikes Peak granite on the eastern slopes of Black Mountain. Colorful dikes that may be of Pikes Peak origin occur in the gneiss and schist south of Bear Creek where there are no outcrops of Pikes Peak granite. Aplitic bodies extend from the Rosalie lobe of the Pikes Peak batholith near the entrance to Harris Park and Deer Creek Park south of the area of Pl. 1.

The small bodies of quartz diorite exposed in Deer Creek and Turkey Creek canyons have little associated pegmatite.

The Silver Plume and Mount Olympus pegmatites of the fifth and last Precambrian cycle carry many genetically related pegmatites. The dikes occur deep in the granite, marginal to it, and in swarms that traverse the injection gneiss enclosing the granite. Pegmatite-rich areas, 1–3 miles wide, border the Indian Creek plutons on the north, northeast, and east and locally cap the granite (Pl. 1). Not every pluton is enclosed by pegmatite-rich gneiss and schist, but each pluton has a partial border of pegmatite. Oval "pillows" of Silver Plume granite are separated by pegmatite that blends into granite without and has a sharp contact with granite within (Pl. 5, fig. 1).

The earliest pegmatite of Silver Plume age filled fractures along the foliation of adjacent schist and gneiss. Later ones of the same cycle invaded sheared zones next to granite. Many such pegmatites trend east-west and dip north, crossing the south-dipping foliation of the Idaho Springs beds.

Aplites are not common diaschistic rocks of Silver Plume and Mount Olympus age. The aplites of the latest generation of granitic rocks are discrete bodies associated with well-zoned pegmatites but are not part of them.

Structural Relations of Pegmatites, Aplites, and Host Rocks

The location of many pegmatites was controlled by structural patterns produced by the emplacement of earlier granitic bodies in highly disturbed bedded rocks, and by the original bedding fissility and partings of the metasedimentaries (Boos and Aberdeen, 1940, p. 727). Uplift of the Front Range, after consolidation of the igneous material of the earlier cycles, opened channels for residual aqueo-igneous material that created richly quartzose pegma-

TABLE 1.—PEGMATITES THAT HAVE BEEN MINED OR PROSPECTED IN THE DENVER MOUNTAIN PARKS AREA, COLORADO

Name and location†	Dimensions and attitude	Orientation of pegmatite and host rocks; contacts	Minerals	Structures	Associated granites	Genesis	Development
SODA CREEK-BEAVER BROOK AREA							
#1. Santa Fe Mtn. Beryl pegmatite, sec. 8, T. 72W. R. 4N. Fig. 2. Altitude about 9200′. Data in part from U.S.G.S. Prof. Paper 227, p. 119–120.	Irregular sheet more than 100′ long, 5′–20′ thick. Depth at least 7′ at E. end. Strike N. 80° W., vertical.	Fills nearly EW. fracture, crosses foliation of orthogneiss that lies NW.-SE. Offshoots follow gneissosity of the gneiss.	Blocky masses of white microcline, irregular bodies of gray quartz. Corroded bluish and greenish beryl crystals associated with albite. Weathered biotite, manganese garnets, and magnetite. Albite replaces beryl up to 6″ across; sericitised albite.	Concentration of rose quartz near middle of dike. No other zones observed. Pegmatite is poor in mica. Beryl crystals angle in from walls of dike and pitch down.	Lies in an area between two granite knobs (U.S.G.S. Prof. Paper 227, p. 30). The granite of this area is the Mount Morrison.	Derived from the end products of igneous activity that produced the gneiss of Santa Fe Mtn.—Saddleback pluton.	Prospected before 1943, nearly a thousand pounds of beryl produced. Open cut about 5′ by 12′ by 1′. Pegmatite exposed to depth of 7′.
#2. Snyder aquamarine pegmatite, sec. 10, T. 4S., R. 72 W. on homesteaded land at about 8000′ elevation on steep S. facing slope. Fig. 2.*	Exposed length about 250′, greatest width 50′. Strike N. 40°-50° E., dip about 70° NW., N. end hooks toward the NE. into the mountain slope.	Dike trends NW. across general NE.-striking foliation of the gneiss. Migmatite on hanging wall side strikes EW., dips N. Main dike paralleled by similar smaller dikes. Footwall contact sharp; dike merges with Mount Morrison migmatite on NW.	Much black quartz, flesh-colored potash feldspar, gray plagioclase, small masses of graphic pegmatite. Small-book silvery muscovite. A little book biotite, rare beryl crystals, aquamarines. Biotite blades fill fractures. Rusty brown garnets. Reddish stains, irregular mass of dark red rock that weathers purplish.	Well-defined zones. Large core area of quartz that is light above and black below. Wall zones are of graphic texture and 2′–4′ thick. Aquamarine embedded in quartz, and near footwall. Brown garnet, quartz, albite in border zones.	Lies in outcrop of Mount Morrison gneiss and injection gneiss. Is in direct contact with parent granitic rock.	Late Mount Morrison pegmatite; aquamarine probably primary and dike developed zone by zone during quiet crystallisation of ichor.	Prospected in 1930; aquamarine mined in 1948 from black quartz core which was trenched 150′ on strike and to 25′ in depth. Operated by C. R. Snyder on whose land the pegmatite occurs.
#3. Beaver Brook School (Brandt) beryl-topaz pegmatite, SW. ¼ sec. 12, T. 4 S., R. 72 W. on north slope of Shaffer Hill, about 7,800′ elevation*. Fig. 2.	Traced 300′ along nearly EW. strike. Middle portion of 150′ strikes N. 80 E. Dip S. 30°–40° into steep slope of Shaffer Hill. Exposed portion about 10′ thick.	Surrounded by and embedded in strongly gneissic Mount Morrison migmatite. Follows orientation of gneissosity of host, which is biotite-rich at contact. Pegmatite blends with Mount Morrison migmatite, locally cuts foliation at low angle.	Green amazonstone, pink to creamy plagioclase, gray to white quartz. Irregularly-spaced small books of muscovite. Topaz, green beryl prisms, minute muscovite flakes. Beryl crystals up to 6″ across in wall zone on footwall side, topaz in small irregular masses. Beryl crystals readily break out of quartz.	Zonation well defined. Core of light green amazonstone (microcline) and gray quartz. Amazonstone extends to footwall. Wall zone contains topaz, muscovite, beryl, and plagioclase. Border zones indistinct, locally blended to adjacent host rock.	Lies within the foliation of Mount Morrison gneiss of Shaffer Hill. Sheets of Boulder Creek granite gneiss near by.	End product of processes that produced the country rock. Some late minerals (tantalite, beryl, topaz) possibly introduced by near by Boulder Creek granite gneiss.	Prospected before 1939, reopened 1944–6. Area thoroughly picked over by local mineral collectors. Shallow trench reveals footwall area of dike.
#4. Grover (Winsor) feldspar, mica, beryl pegmatite, chiefly in sec. 9, T. 4 S., R. 72 W., E. end of Santa Fe Mtn. at about 8730′ elevation.* Figs. 2, 3.	Tabular body, thickest at E.; exposed 750′ on strike of EW., forks at W. end; nearly vertical. Walls nearly 40′ apart in thickest part. Hanging wall dips SE. about 70°. Plunges E. at low angle. Mount Morrison migmatite drag-folded against pegmatite in hanging wall.	Enclosed in granitised Idaho Springs beds and Mount Morrison migmatite. Gneissosity of host strikes 60°–70° on NW. contact of pegmatite. Pegmatite crosses host at low angle on NE. Buranek and Crawford: "Bertrandite, beryl and other economic minerals of Grover Pegmatite", Vol. 21, Proc. Utah Acad. Sci., 1943–1944, p. 103.	Gray to white quartz abundant, white to creamy albite, green potash feldspar. Amazonstone disposed of in early days as beryl*. Books of muscovite with clear sheets. Magnetite, beryl, biotite, 2 generations of muscovite, garnet, schorl, columbite; bertrandite, rutile, fluorite, monazite, tantalite. Malachite stains in feldspar cleavage and fractures in quartz. Quartz cap at W. stained red. * S. A. Ionides, oral communication August, 1939.	Well-defined zones. Core of quartz extends length of the dike and to depth of at least 30′. Wall zones of albite, biotite, quartz, microcline; intermediate zones have albite, muscovite, beryl, fluorite. Narrow border zones at both contacts except where blended with Mount Morrison gneiss.	Dike in direct contact with a small body of Mount Morrison gneiss on southern margin. No other granitic rock within 1000′. Follows structural lines of the Idaho Springs beds into which it was intruded.	Pegmatite confined to limited channel; probable end product of granitising processes that produced the injection gneiss. Beryl, muscovite, albite, and tantalite fed into the intermediate zones by replacing solutions. Aggressive movement of ichor dragged the gneiss of the host rock.	Opened as feldspar mine, 1937. The owner, Joe Grover, produced feldspar, mica and beryl from 1940–1945. Maximum opening a 2-level trench about 40′ deep at E., 300′ long and 25′ wide. Western half of dike prospected by pits.
#5. Coors feldspar pegmatite, E. center sec. 18, T. 4 S., R. 71 W. at NW. end of ridge running N. from Bergen Park. Elevation about 8000′. Figs. 2, 6*.	Two lensing pegmatites *en echelon*. East dike strikes N-NW., dips S. Dikes occupy part of a local syncline about 550′ long and 250′ across. Axial strike N. 60° E. Pegmatites very irregular in shape, pinch and swell along walls. Both dikes thicken and blunt out to W. and SW. and thin out to NE.	Host rock chiefly hornblende gneiss and lime-silicate and carbonate beds, foliated NW. and dipping into syncline. Pegmatite intruded nearly concordantly with bedding of the gneiss. The two pegmatite bodies cross the regional NW. trend of the metasedimentaries.	Cleavelandite, potash feldspar, white microcline, white to smoky glistening quartz, graphic pegmatite, frondose and burr silvery muscovite, small amounts of rose quartz, rare books of muscovite. Streaks and small ribs of magnetic iron oxides, brown earthy garnets, rare blades of biotite, orthite; uranium-yellow stains in quartz.	W. dike: nearly continuous quartz core pinches and swells forming hood at the W. and SW. Wall zones graphic and with white microcline and plagioclase. White feldspar in both dikes. Good feldspar below quartz hood. Festoons and sunbursts of silvery muscovite in feldspar of wall zones. Quartz core of SE. dike nearer footwall than hanging wall and rakes to SW. Conspicuous narrow border zones in places.	Closely associated with small bodies of Mount Morrison injection gneiss and migmatite. One dike seems to rise from Mount Morrison gneiss directly below it. Dike grades into Mount Morrison gneiss at NE.	Original dikes were simple pegmatites consisting of potash feldspar surrounding simple quartz core. Cleavelandite replaced microcline, and muscovite came into cleavages of microcline. Black quartz introduced into the areas of white quartz.	Operated for feldspar and scrap mica by Coors Pottery Co. of Denver. Early openings before 1937. Shafts and short tunnels caved. Trenches open, limited to gophering around quartz core bodies.
#6. East Shaffer Hill pegmatites, SW. ¼ sec. 7, T. 4 S., R. 71 W., on ridge E. of road crossing Shaffer Hill at about 7700′ elev.* Figs. 2, 5.	Two interconnected dikes on S. slope of ridge, N. 70° E. to EW. trends. Exposed for more than 800′ on strike and 300′ N-S. Main dike 20′ thick, much warped and branched. Pegmatites appear to lie in syncline and outcrop on limbs.	Outcrops lie nearly at right angles to regional NW. trends of Mount Morrison gneiss and Swandyke gneiss host. Footwall rolls and pinches out against the gneiss where dips are steep. Contacts well defined. Discontinuous chill margins in the pegmatites.	Quartz rosy to bluish white: microcline, pencil and cuneiform graphic pegmatite, lumps of black to gray quartz in potash feldspar. Small amount of cleavelandite; brown garnets in small clusters, a little magnetite, some schorl. Poor in micas. Tawny microlite stains on feldspar; green malachite stains in feldspar cleavages; greenish to rusty sheets in biotite.	Well-defined rose to white or bluish quartz and potash feldspar core. Patchy graphic and potash feldspar wall zones make sheath around core. A little biotite in all zones. Border zones are persistent, narrow aggregates of fine-grained quartz and plagioclase.	A sheet of Mount Morrison gneiss interlayered with hornblende gneiss and injection gneiss.	Sheets of pegmatite intruded into metamorphic bedded rocks and folded with them into an asymetrical syncline.	Prospected about 1938; limited mining about 1945 produced feldspar. Three open cuts, deepest about 8′. Feldspar gophered out along quartz bodies. Quarts core must be moved to reach more feldspar.
#7. Old Soda Creek School feldspar pegmatite, N.W. ¼, sec. 18, T. 4 S., R. 71 W., on south side of Soda Creek valley, 50′ above water level. Fig. 2*.	Strike EW. to N. 60° W.; thick and blunt at W. end. Appears to widen at depth, N. wall rolls out into host rock. Oval pod, maximum thickness of 300′. Opened to 30′ depth; nearly vertical.	Host rock on N. a granitised dioritic gneiss, foliation N. 60° W. encloses pegmatite on NE. Pegmatite crosses foliation of hornblende gneiss and lime-silicate beds on its south wall; N. wall rolled vertically.	Rose, bluish, and white quartz present in various parts of dike. Cleavelandite locally abundant; graphic pegmatite and microcline; books of rusty biotite fill fractures. Streaks of small brown garnets, blades of tantalite, minute grains of microlite, no muscovite. Yellow stains in quartz, malachite stains on micas.	Quartz core discontinuous and of irregular shape, rosy below and white above; pearly cleavelandite under quartz of core. Discontinuous graphic potash feldspar wall zones. Biotite in all zones. Border zones persistent narrow aggregates of fine-grained albite and quartz.	Sheets of a dioritic rock, similar to the Boulder Creek gneiss, and pods of Mount Morrison gneiss are adjacent to N. side of the pegmatite.	Lensing body of pegmatite originally of late Mount Morrison genesis, modified by Boulder Creek magma, coincident with the development of Boulder Creek gneiss near by.	Opened on two levels, large quartz core exposed 1939. In 1941, upper part of dike stripped off, at least 25′ of dike depth exposed. Bottom of pegmatite not reached in mining.
#8. Feldspar pegmatite, SE. ¼ sec. 7, T. 4 S., R. 71 W., on SW.-facing slope. Fig. 2.*	Thick pod of pegmatite elongated to NE. Strike N. 20° E. into steep slope of hillside. Exposed to depth of 25′. Strike length not known; vertical to dipping steeply SE. Thickness up to 20′; walls roll slightly.	Enclosed by massive, greenish hornblendic gneiss, with local porphyroblastic layers. Pegmatite crosses the N. 30°–50° E. foliation of gneiss.	Black and bluish-white quartz, flesh-colored microcline, tan to creamy plagioclase. Cuneiform graphic pegmatite. Rusty to golden biotite fills fractures in quartz and in feldspars and in sheared places in pegmatite. Soft black tryphyllite. Rusty garnets. Unknown powdery black mineral in pockets. Yellow stains in quartz, possibly uranium.	Large nearly vertical quartz core, 6′–8′ thick, pinches and swells. Wall zones of graphic and perthitic feldspar, bluish gray quartz, and dull white plagioclase. Isolated garnet dodecahedrons. Rusty golden biotite. No border zones.	No granitic rock crops out on this slope. Doubtful Silver Plume granite about ½ mile to east of pegmatite.	Closely associated with dikes of hornblende gneiss but pegmatite has aspects of a granitic origin.	Mined intermittently since 1947.
SWEDE GULCH-KERR GULCH-KITTREDGE AREA							
#9. Sunrise Peak quartz and feldspar pegmatite, SW. ¼ sec. 22, T. 4 S., R. 71 W. on E. side of Kerr Gulch at about 7700′ elevation. Known locally as Traut-Rudin-Anderson pegmatite. Figs. 2, 7.*	Outcrop length 800′ EW. Dip N. 80° into Sunrise Peak. Rakes down to E. Up to 50′ thick at W., narrows to 10′ at E. end. Strike of tabular body N. 60–70° E. where opened. Dip. SE. on hanging wall. Vertically exposed for 40′, probably thickens with depth.	EW.-foliated hornblende gneiss at N. contact; pegmatite crosses and locally blends with Mount Morrison migmatite on S. and SE. Large siliceous W. end terminates against sheared, massive hornblende gneiss. Dike crosses crumpled lime-silicate beds that strike N. 70° W. Dike sharply set off from host rocks and not influenced by their character. Two - feet - wide shear zone with broken and stained rock against dike on SE.	White, rose and smokey quartz at W. end of dike. Flesh-colored euhedral potash feldspar under quartz. Plagioclase in E. end of dike. Some 525′ of quarts drilled in 1940, white above and smoky below. Euhedral books of fresh black biotite, wedges of rusty biotite in quarts, schorl prisms and veins, beryl, samarskite, tryphyllite, tantalite nodules, thin plates of reddish garnets, rusty garnet, topaz. Specular hematite in joints, soft black unknown mineral with biotite. Sheets of white sericite in quarts of W. end of dike.	Zonation entire length of dike. Fine-grained border 3′–5′ thick on N. wall. Core of rose and milky to gray quarts, underlain at W. end by large crystals of microcline. Wide wall zone of graphic albite with frondose quarts on foot wall side. W. end sheathed in aplitic granite. Muscovite books where dike pinches down and in crest of pegmatite. Ill-defined border zones.	Eastern part of pegmatite in contact with gneissic Mount Morrison. A lense of Silver Plume granite lies on south slope of Sunrise Peak, about 100′ vertically above the pegmatite.	Probably of late Mount Morrison age, modified by ichor that developed the Silver Plume bodies nearby. Intermediate zones of dike appear to be replacement units.	West end opened about 1940 and Brook Forest Inn built of quartz. Quartz was crushed and applied to lighten surface of 17th street in Denver. Feldspar produced from center portion 1925–1942. Open cut and drift to 40′, 1945. Partly filled with water 1946–1950.
#10. Silver Glen Ranch mica-feldspar pegmatite, SW. ¼ sec. 26, T. 4	Tabular pegmatite exposed for 400′ on strike NW.	Pegmatite crosses foliation of hornblende gneiss,	Varicolored green, milky, smoky or faintly rosy, glistening quartz in lobate bodies	Well-zoned with thick quarts core at least 35′ through; lobate margin. Creamy	Closely associated with sheets of Mount Morrison migma-	This strongly micaceous pegmatite was developed pri-	Opened about 1945 and mined for feldspar and mica. No

#7. Old Soda Creek School feldspar pegmatite, N.W. ¼, sec. 18, T. 4 S., R. 71 W., on south side of Soda Creek valley, 50' above water level. Fig. 2*. * Examined by M. F. Boos, June, 1939 and Dec. 13, 1950.	Strike EW. to N. 60° W.; thick and blunt at W. end. Appears to widen at depth, N. wall rolls out into host rock. Oval pod, maximum thickness of 500'. Opened to 30' depth; nearly vertical.	Host rock on N. a granitized dioritic gneiss, foliation N. 60° W. encloses pegmatite on NE. Pegmatite crosses foliation of hornblende gneiss and lime-silicate beds on its south wall; N. wall rolled vertically.	Rose, bluish, and white quartz present in various parts of dike. Cleavelandite locally abundant; graphic pegmatite and microcline; books of rusty biotite fill fractures. Streaks of small brown garnets, blades of tantalite, minute grains of microlite, no muscovite. Yellow stains in quartz, malachite stains on micas.	Quartz core discontinuous and of irregular shape, rosy below and white above; pearly cleavelandite under quarts of core. Discontinuous graphic potash feldspar wall zones. Biotite in all zones. Border zones persistent narrow aggregates of fine-grained albite and quartz.	Sheets of a dioritic rock, similar to the Boulder Creek gneiss, and pods of Mount Morrison gneiss are adjacent to N. side of the pegmatite.	Lensing body of pegmatite originally of late Mount Morrison genesis, modified by Boulder Creek magma, coincident with the development of Boulder Creek gneiss near by.	Opened on two levels, large quarts core exposed 1939. In 1941, upper part of dike stripped off, at least 25' of dike depth exposed. Bottom of pegmatite not reached in mining.
#8. Feldspar pegmatite, SE. ¼ sec. 7, T. 4 S., R. 71 W., on SW.-facing slope. Fig. 2.* * Examined by M. F. Boos, Dec. 13, 1950.	Thick pod of pegmatite elongated to NE. Strike N. 20° E. into steep slope of hillside. Exposed to depth of 25'. Strike length not known; vertical to dipping steeply SE. Thickness up to 20'; walls roll slightly.	Enclosed by massive, greenish hornblendic gneiss, with local porphyroblastic layers. Pegmatite crosses the N. 30°–50° E. foliation of gneiss.	Black and bluish-white quartz, flesh-colored microcline, tan to creamy plagioclase. Cuneiform graphic pegmatite. Rusty to golden biotite fills fractures in quartz and in feldspars and in sheared places in pegmatite. Soft black tryphyllite. Rusty garnets. Unknown powdery black mineral in pockets. Yellow stains in quartz, possibly uranium.	Large nearly vertical quartz core, 6'–8' thick, pinches and swells. Wall zones of graphic and perthitic feldspar, bluish gray quartz, and dull white plagioclase. Isolated garnet dodecahedrons. Rusty golden biotite. No border zones.	No granitic rock crops out on this slope. Doubtful Silver Plume granite about ½ mile to east of pegmatite.	Closely associated with dikes of hornblende gneiss but pegmatite has aspects of a granitic origin.	Mined intermittently since 1947.
SWEDE GULCH-KERR GULCH-KITTREDGE AREA **#9.** Sunrise Peak quartz and feldspar pegmatite, SW. ¼ sec. 22, T. 4 S., R. 71 W. on E. side of Kerr Gulch at about 7700' elevation. Known locally as Traut-Rudin-Anderson pegmatite. Figs. 2, 7.* * Examined by M. F. Boos, June 17, 1940, July 11, 1941, Nov. 4, 1945. Mapped July 4, 1946.	Outcrop length 800' EW. Dip N. 80° into Sunrise Peak. Rakes down to E. Up to 50' thick at W., narrows to 10' at E. end. Strike of tabular body N. 60–70° E. where opened. Dip. SE. on hanging wall. Vertically exposed for 40', probably thickens with depth.	EW.-foliated hornblende gneiss at N. contact; pegmatite crosses and locally blends with Mount Morrison migmatite on S. and SE. Large siliceous W. end terminates against sheared, massive hornblende gneiss. Dike crosses crumpled lime-silicate beds that strike N. 70° W. Dike sharply set off from host rocks and not influenced by their character. Two - feet - wide shear zone with broken and stained rock against dike on SE.	White, rose and smokey quartz at W. end of dike. Flesh-colored euhedral potash feldspar under quartz. Plagioclase in E. end of dike. Some 525' of quarts drilled in 1940, white above and smoky below. Euhedral books of fresh black biotite, wedges of rusty biotite in quarts, schorl prisms and veins, beryl, samarskite, tryphyllite, tantalite nodules, thin plates of reddish garnets, rusty garnet, topaz. Specular hematite in joints, soft black unknown mineral with biotite. Sheets of white sericite in quartz of W. end of dike.	Zonation entire length of dike. Fine-grained border 3'–5' thick on N. wall. Core of rose and milky to gray quartz, underlain at W. end by large crystals of microcline. Wide wall zone of graphic albite with frondose quarts on foot wall side. W. end sheathed in aplitic granite. Muscovite books where dike pinches down and in crest of pegmatite. Ill-defined border zones.	Eastern part of pegmatite in contact with gneissic Mount Morrison. A lense of Silver Plume granite lies on south slope of Sunrise Peak, about 100' vertically above the pegmatite.	Probably of late Mount Morrison age, modified by ichor that developed the Silver Plume bodies nearby. Intermediate zones of dike appear to be replacement units.	West end opened about 1940 and Brook Forest Inn built of quarts. Quarts was crushed and applied to lighten surface of 17th street in Denver. Feldspar produced from center portion 1925–1942. Open cut and drift to 40', 1945. Partly filled with water 1946–1950.
#10. Silver Glen Ranch mica-feldspar pegmatite, SW. ¼ sec. 26, T. 4 S., R. 71 W. on SE. slope of hill at about 7200' elev. and 200' above stream in Swede Gulch. Fig. 2.* * Examined by M. F. Boos, June 12, 1951.	Tabular pegmatite exposed for 400' on strike NW. 60°-70°, vertical. Dike dips SW. up to 80° SE. end cut off by erosion. NW. end rakes into hillside. Exposed to depth of 25' and appears to thicken downwards.	Pegmatite crosses foliation of hornblende gneiss, quarts gneiss, and injected Mount Morrison gneiss. NE. wall sheared against Mount Morrison. Body of dike generally follows foliation of host rock. Faulted eastern contact.	Varicolored green, milky, smoky or faintly rosy, glistening quarts in lobate bodies with irregular fractures, creamy microcline streaked with pale flesh colored feldspar. Dull white cleavelandite with quarts. Numerous festoons of silvery herringbone, interlocked, muscovite books. No biotite. Rusty brown garnets, topaz, microlite, burr muscovite, white beryl.	Well-zoned with thick quarts core at least 35' through; lobate margin. Creamy microcline and graphic feldspar surround the core. Dull plagioclase and muscovite in wall zones. Festoons of herringbone mica fan out into quartz core from wall zones. Little or no border zone exposed. Dike crossed by several N. 65° W. faults, movement horizontal.	Closely associated with sheets of Mount Morrison migmatite. Several small lenses of Silver Plume granite near NW. end of dike.	This strongly micaceous pegmatite was developed primarily in micaceous gneiss and schist by the agents that produced the Mount Morrison migmatite. Later, during Silver Plume times, it was modified by the introduction of muscovite and accessory minerals.	Opened about 1945 and mined for feldspar and mica. No mining since 1947.
#11. Seven Hills lode feldspar-mica pegmatite S. ½ sec. 34, T. 4 S., R. 71 W. Locally known as "Yetter" pegmatite; at stream level on S.-facing slope. Fig. 2.* * Visited by M. F. Boos, June 1940, June 1951.	Strike nearly EW., dip 70° NE. N. wall of dike dips NW. 80°, strike N. 60° E. Length about 250', thickness 25'. S. end cut off by erosion.	Enclosed by injection gneiss and biotite-sillimanite schist of Idaho Springs beds, foliation nearly EW. Pod of Silver Plume granite on north wall of pegmatite.	White to gray quartz, light flesh-colored microcline, numerous small, thick, clean books of silvery muscovite; biotite in thin strips filling cleavages and fractures. Some larger books of muscovite carry clear mica. Schorl and small garnets.	Primarily a distinctly-zoned dike, with white to grayish quarts core, wall zones of light pink microcline thickly studded with small books of clear muscovite. Biotite blades in fractures. Border zone about 6" thick in N. wall.	Lense of Silver Plume granite attached to N. wall of dike.	Probably developed during late Silver Plume times.	More than half the dike mined out to obtain ornamental rock for numerous buildings on the Yetter Ranch.
#12. Swede Gulch pegmatites, secs. 26, 27, 35, 36, T. 4 S, R. 71 W. Fig. 2.* * Examined by M. F. Boos, June 13, 1940, July 11, 1941, July 7, 1941, Aug. 16, 1943.							
Dike on E. side of Swede Gulch, NW. cor. sec. 26.	Lens 100' long and 20' thick. Strike NW., dips steeply NE. into slope.	Concordant with and enclosed by bedded hornblende gneiss that is banded and foliated N. 60° W., dip 60° E.	Microcline, cleavelandite, milky and smoky quarts, sunbursts of muscovite in microcline. Biotite intergrown with muscovite. Rusty garnets.	Coarse-textured core of quarts and microcline. Sunbursts of muscovite border the core. Footwall has border of white graphic pegmatite. Hanging wall carries narrow, combined wall- and border-zones.	Small sheets of both Mount Morrison migmatite and Boulder Creek gneiss outcrop within 500' of dike. Also a pod of questionable Silver Plume granite.	Probably of mixed origin; originated by magma that created the sheets of Mount Morrison and replaced by Silver Plume on a limited scale.	Opened by a prospect pit before 1940.
Dike near valley floor on W. side of Swede Gulch, NW. ¼ sec. 26.	200' long on strike of N. 60° W.; vertical to NE. dip. 25' thick. One of 6 subparallel sills trending NW.	*Lit-par-lit* with the foliation of hornblende gneiss and mica schist that surround the dike. Body of dike lensing against gneiss.	Microcline, graphic pegmatite, massive quarts of gray color. Burr muscovite, garnets.	Ill-defined quarts-microcline core area. Wall zones of graphic pegmatite.	Numerous sills of Mount Morrison migmatite in near by gneiss.	Probably a simple early Mount Morrison pegmatite, developed by magmatic invasion of the host.	Worked for feldspar before 1942.
Top of ridge, NE. corner sec. 27 at about 7500' elevation.	Sprawling, branched dike nearly horizontal. Outcrops 200' across. Dike 30'–50' thick.	Crosses NW.-SE. foliation of gneiss and schist on the N. contact and is against Mount Morrison migmatite on S. wall.	Milky white quarts, microcline, small amount of creamy plagioclase. Much small-size burr mica.	Muscovite books in thin layers of quarts, near middle of outcrop, bordered by graphic pegmatite.	Crosses the foliation of Mount Morrison gneiss and may be associated with Boulder Creek gneiss at depth.	Probably a late-forming Mount Morrison pegmatite, replaced in the core area by salmon-colored feldspar of Boulder Creek age.	Prospected by trenching in 1943.
E. side of junction of Swede Gulch with Bear Creek.	Sill 100' long and 10'–15' thick. Trends N. 30°–60° W., dips steeply NE.	Concordant with NW.-SE. foliation of hornblende gneiss and schist; grades into Mount Morrison gneiss on N.	White quarts, units of coarse flesh-colored microcline, minor cleavelandite. Small books of clear muscovite. Beryl.	Core of quarts and potash feldspar exposed near center of area prospected. Wall zone chiefly of graphic pegmatite.	Closely associated on the outcrop with fine-grained Mount Morrison gneiss.	Probably of Mount Morrison age and development.	Small prospect pit in S. part of dike, 1940.
#13. Wasson (Rainbow Hills) beryl-feldspar pegmatite, sec. 15, T. 4 S., R. 71 W. Lies on crest of low ridge, Figs. 2, 8.* * Examined by M. F. Boos, June, 1939, July 9, 1940, Aug. 16, 1945. Mapped Aug. 17, 1946, Oct. 1949.	Traced for 650', is convex to NE. Hanging wall dips SW. 70°–80°. Body of dike dips SW., and pinches out to the NW. and SE. Greatest width 50', depth 75'.	Host rock a foliated diorite, Boulder Creek(?) age. Gneissosity of diorite N. 50° W., dip 60° NE. Pegmatite crosses foliation in middle area but conforms to it at NW. end. Diorite cut by aplitic granite probably Silver Plume age.	Rose and white quarts, smoky quarts embedded in white quarts, small amount of plagioclase, muscovite and a very little biotite. Graphic pegmatite at E. end of dike, and dark gray to amethystine quarts in feldspar. Small book muscovite in quarts. Thin rhombs of biotite in muscovite; rusty biotite; green beryl prisms in branch of main pegmatite. Rusty brown garnets in clusters with a dark earth.	Quartz-microcline core forms a low ridge. Core in hanging wall area at E. end of dike. Six foot wall zones of amethyst quarts, microcline and some plagioclase. No border zones exposed. Intermediate zones discontinuous. Core at least 20' deep, occuping a low, easily-traced ridge. Beryl in pockets in main dike and in small branch dike.	Pegmatite crosses outcrop of Mount Morrison and blends with it. Isolated bodies of Silver Plume granite a short distance to N.	Late Mount Morrison dike developed by ichor. May have been replaced in part by Silver Plume magma.	Prospected for beryl about 1938; small pit in branch of main dike, E. part opened and mined for feldspar about 1944.
SWEITZER GULCH-TWIN FORKS AREA **#14.** Bigger mica-beryl pegmatite, Sec. 3, T. 6 S., R. 70 W., near N. end of Bald Mtn. Locally known as the Sweitzer lode. Figs. 2, 4, Pl. 5.* * Examined by M. F. Boos, Nov. 11, 1939; July 13, 1939; Aug. 13, 1941; mapped Nov. 4, 1945; visited July, 1951, June 1953.	Traced on outcrop for 650'; trend convex to NE. Hanging wall dips SW. 70°–80°. Body of dike dips SW. to vertical. Pegmatite pinches out to the NW. and SE. but appears to grade into sheet of Silver Plume aplitic granite. Greatest width about 50', depth 75'.	Host rock chiefly foliated diorite of probable Boulder Creek age. Gneissosity N. 50° W., dip NE. 60°. Pegmatite cuts foliation at high angle in middle of outcrop, but conforms at NW.	Massive blocks of flesh-colored potash feldspar, graphic pegmatite, white to smoky quarts, rare plagioclase, chiefly albite, light rose quarts, books of muscovite, rare books of biotite. Beryl logs, schorl, garnets, bismuthite, bismuthinite, pockets of sulphides, monazite, columbite. Kaolinized feldspars, albitized beryl; soft, white, earthy unknown mineral adjacent to the bismuthinite.	Dike is well-zoned in length, depth and on the dip. Microcline, rose to white quarts core 10'–15' thick locally extending from wall to wall. Foot wall area of graphic pegmatite, hanging wall of biotite, muscovite froods and festoons and gray quarts and microcline. White albite and schorl adjacent to core. Intermediate zones well developed with beryl, albite, cleavelandite, tantalite, and schorl.	Adjacent to and partly enclosed by dioritic Boulder Creek gneiss-part of Fairy Hill pluton. Silver Plume dikes cross the Boulder Creek, and Silver Plume lies adjacent to the pegmatite at its S. end.	Complex dike initiated in Boulder Creek times, suffered replacement and alteration by the residues of Silver Plume magma.	Mined for mica 1885–1900. Reopened about 1937 for feldspar. Quarts core removed about 1939, pipes of muscovite on W. wall exposed and mined. Beryl logs removed about 1942.
#15. Thomas mica pegmatite, NW ¼ sec. 11, T. 6 S., R. 70 W. on W. slope of Mica/ Mine Creek at about 7,000' elev.* Fig. 2. * Visited by M. F. Boos, July 6, 1940. Information also from U.S.G.S. Bull. 740, p. 58	Irregular dike trends N. 60° W. Steep walls. NW. end hidden in slope of valley wall.	Host rock chiefly a dioritic phase of Boulder Creek gneiss, interlayered with fissile mica schist. Gneissosity N. 50° W., dip 60° NE. Dike pinches down to SE. Micaceous parts of dike where coa-	Even-grained, fine to coarse potash feldspar; quarts, scrap-type muscovite and some good sheet. Apatite and small red garnets.	Rather even, fine grain of pegmatite is spotted with coarse quarts-feldspar units, and with bunches of book muscovite and festoons of froodose muscovite. No core seen.	Dioritic gneiss, much like Boulder Creek gneiss of the Bigger pegmatite area is adjacent to the dike. Several sheets of Mount Morrison migmatite cross to north of dike.	Pegmatite probably developed in late Boulder Creek times. Reddish granite may be Silver Plume.	Area littered with scrap mica. Mined before 1923 by open cut and drift.

Top of ridge, NE. corner sec. 27 at about 7500′ elevation.	Spreading, irregular and nearly horizontal. Outcrops 200′ across. Dike 30′–50′ thick.	of gneiss and schist on the N. contact and is against Mount Morrison migmatite on S. wall.	amount of creamy plagioclase. Much small-size burr mica.	near middle of outcrop, bordered by graphic pegmatite.	associated with Boulder Creek gneiss at depth.	in the core area by salmon-colored feldspar of Boulder Creek age.	
E. side of junction of Swede Gulch with Bear Creek.	Sill 100′ long and 10′–15′ thick. Trends N. 50°–60° W., dips steeply NE.	Concordant with NW.-SE. foliation of hornblende gneiss and schist; grades into Mount Morrison gneiss on N.	White quartz, units of coarse flesh-colored microcline, minor cleavelandite. Small books of clear muscovite. Beryl.	Core of quartz and potash feldspar exposed near center of area prospected. Wall zone chiefly of graphic pegmatite.	Closely associated on the outcrop with fine-grained Mount Morrison gneiss.	Probably of Mount Morrison age and development.	Small prospect pit in S. part of dike, 1940.
#13. Wasson (Rainbow Hills) beryl-feldspar pegmatite, sec. 15, T. 4 S., R. 71 W. Lies on crest of low ridge, Figs. 2, 8.*	Traced for 650′, is convex to NE. Hanging wall dips SW. 70°–80°. Body of dike dips SW., and pinches out to the NW. and SE. Greatest width 50′, depth 75′.	Host rock a foliated diorite, Boulder Creek(?) age. Gneissosity of diorite N. 50° W., dip 60° NE. Pegmatite crosses foliation in middle area but conforms to it at NW. end. Diorite cut by aplitic granite probably Silver Plume age.	Rose and white quartz, smoky quartz embedded in white quartz, small amount of plagioclase, muscovite and a very little biotite. Graphic pegmatite at E. end of dike, and dark gray to amethystine quartz in feldspar. Small book muscovite in quartz. Thin rhombs of biotite in muscovite; rusty biotite; green beryl prisms in branch of main pegmatite. Rusty brown garnets in clusters with a dark earth.	Quartz-microcline core forms a low ridge. Core in hanging wall area at E. end of dike. Six foot wall zones of amethyst quartz, microcline and some plagioclase. No border zones exposed. Intermediate zones discontinuous. Core at least 20′ deep, occuping a low, easily-traced ridge. Beryl in pockets in main dike and in small branch dike.	Pegmatite crosses outcrop of Mount Morrison and blends with it. Isolated bodies of Silver Plume granite a short distance to N.	Late Mount Morrison dike developed by ichor. May have been replaced in part by Silver Plume magma.	Prospected for beryl about 1938; small pit in branch of main dike, E. part opened and mined for feldspar about 1944.
* Examined by M. F. Boos, June, 1939, July 9, 1940, Aug. 16, 1945. Mapped Aug. 17, 1946, Oct. 1949.							
SWEITZER GULCH-TWIN FORKS AREA							
#14. Bigger mica-beryl pegmatite, Sec. 3, T. 6 S., R. 70 W., near N. end of Bald Mtn. Locally known as the Sweitzer lode. Figs. 2, 4, Pl. 5.*	Traced on outcrop for 650′; trend convex to NE. Hanging wall dips SW. 70°–80°. Body of dike dips SW. to vertical. Pegmatite pinches out to the NW. and SE. but appears to grade into sheet of Silver Plume aplitic granite. Greatest width about 50′, depth 75′.	Host rock chiefly foliated diorite of probable Boulder Creek age. Gneissosity N. 50° W., dip NE. 60°. Pegmatite cuts foliation at high angle in middle of outcrop, but conforms at NW.	Massive blocks of flesh-colored potash feldspar, graphic pegmatite, white to smoky quartz, rare plagioclase, chiefly albite, light rose quartz, books of muscovite, rare books of biotite. Beryl logs, schorl, garnets, bismuthite, bismuthinite, pockets of sulphides, monazite, columbite. Kaolinized feldspars, albitised beryl; soft, white, earthy unknown mineral adjacent to the bismuthinite.	Dike is well-zoned in length, depth and on the dip. Microcline, rose to white quartz core 10′–15′ thick locally extending from wall to wall. Foot wall area of graphic pegmatite, hanging wall of biotite, muscovite fronds and festoons and gray quartz and microcline. White albite and schorl adjacent to core. Intermediate zones well developed with beryl, albite, cleavelandite, tantalite, and schorl.	Adjacent to and partly enclosed by dioritic Boulder Creek gneiss-part of Fairy Hill pluton. Silver Plume dikes cross the Boulder Creek, and Silver Plume lies adjacent to the pegmatite at its S. end.	Complex dike initiated in Boulder Creek times, suffered replacement and alteration by the residues of Silver Plume magma.	Mined for mica 1888–1890. Reopened about 1937 for feldspar. Quarts core removed about 1939, pipes of muscovite on W. wall exposed and mined. Beryl logs removed about 1942.
* Examined by M. F. Boos, Nov. 11, 1939; July 13, 1939; Aug. 13, 1941; mapped Nov. 4, 1945; visited July, 1951, June 1953.							
#15. Thomas mica pegmatite, NW ¼ sec. 11, T. 6 S., R. 70 W. on W. slope of Mica Mine Creek at about 7,000′ elev.* Fig. 2.	Irregular dike trends N. 60° W. Steep walls. NW. end hidden in slope of valley wall.	Host rock chiefly a dioritic phase of Boulder Creek gneiss, interlayered with fissile mica schist. Gneissosity N. 50° W., dip 60° NE. Dike pinches down to SE. Micaceous parts of dike where pegmatite contacts the mica schist.	Even-grained, fine to coarse potash feldspar; quartz, scrap-type muscovite and some good sheet. Apatite and small red garnets.	Rather even, fine grain of pegmatite is spotted with coarse quartz-feldspar units, and with bunches of book muscovite and festoons of frondose muscovite. No core seen.	Dioritic gneiss, much like Boulder Creek gneiss of the Bigger pegmatite area is adjacent to the dike. Several sheets of Mount Morrison migmatite cross to north of dike.	Pegmatite probably developed in late Boulder Creek times. Reddish granite may be Silver Plume.	Area littered with scrap mica. Mined before 1923 by open cut and drift.
* Visited by M. F. Boos, July 6, 1940. Information also from U.S.G.S. Bull. 740, p. 58							
#16. Fenders mica pegmatite near C., sec. 34, T. 5 S., R. 70 W. at about 7,000′ elev. on E. side Turkey Creek. Fig. 2.*	Trends NW.-SE., body of dike nearly horizontal in places. Limits embedded in slope of hill. Pegmatite flat on base, rolled and convex on top. Maximum length 1000′.	Crosses foliation of schist and gneiss. W. edge of dike cuts body of hornblende gneiss. Mica schist on top of dike coarsened by contact. Base of dike sheared against host rock.	White and rose quartz, potash feldspar, small amount of plagioclase, sunbursts of muscovite and some poor sheet. Rusty garnets, bladed biotite, schorl, olive green mineral in muscovite. Soft, unknown, powdery, black to rusty mineral.	Zones poorly defined. White quartz underlain by burr mica and garnet. Good concentrations of potash feldspar near lower edge of dike. No border zones exposed.	Enclosed in Mount Morrison migmatite and also closely associated with pods of Boulder Creek gneiss.	Pegmatite initiated in Mount Morrison times, recrystallized and replaced by invasion of Boulder Creek ichor. Part of original coarse-textured core destroyed by later mineralization.	Pegmatite prospected before 1940. Western face of outcrop exposed to depth of 20′.
* Examined by M. F. Boos, Aug. 13, 1941.							
#17. Iowa Gulch feldspar-mica pegmatite, sec. 35, T. 5 S., R. 70 W. N. slope of Iowa Creek valley. Fig. 2.*	Flat dike covering about ¼ square mile. Branches and forks. One of a swarm of pegmatites in the area.	Host rocks mica schist and injection gneiss foliated N. 50° E., dip NW. steeply. Dike crosses trends in host rock. Pegmatite elongated NW.-SE., dip NW. Mica gneiss and schist silicified at contact.	White and rosy quartz, microcline, a little plagioclase. Small books of silvery muscovite. Biotite in thin books and strips.	Definitely zoned, with quartz core cap underlain by potash feldspar, mixed gray quartz, small muscovite books. Wall zones of albite and muscovite; quartz poorly exposed.	Lenticular outcrops of Silver Plume granite near by. No other granite within ½ mile.	Appearance, mineral content, and structure of a Silver Plume pegmatite.	Prospect pits in N. face of ridge have opened dike to 10′ deep.
* Examined by M. F. Boos, Aug. 15, 1949.							
#18. Medlen School feldspar pegmatite, SE. ¼ sec. 33, T. 5 S., R. 70 W., on E.-facing slope at about 7300′ elevation. Fig. 2.*	Complex area with several NW.-SE.-trending pegmatites. Main dike lies NS., vertical; 150′ long and 20′ thick.	Host rock fine-grained mica gneiss and schist. Dense, tough hornblende gneiss exposed on E. side and mica schist on S.	Sunbursts of muscovite, clusters of small muscovite books, rare books up to 12″ across but reeved and broken. Masses of flesh-colored potash feldspar. Rusty brown garnets.	Quartz core lacking. Dike is 90 percent feldspar. Festoons of muscovite in potash feldspar. Gray quartz bodies have rare books of muscovite. Narrow border zones of muscovite, albite, quartz.	Dike a mixture of granite and pegmatite. Exposed on W. margin of Fairy Hill stock of Boulder Creek gneiss. Outcrops of lensing Silver Plume granite invade dike from the W.	Boulder Creek pegmatite modified by Silver Plume magma; silvery muscovite introduced and dark gray to black quartz.	Opened, and prospected with trenches about 1944.
* Examined by M. F. Boos, Sept. 17, 1949.							
#19. Mount Judge beryl pegmatites, sec. 28, T. 4 S., R 72 W., on NE. end of Judge and W. of old sawmill. Fig. 2.*	Swarm of small dikes trending NE. Some 100′-150′ long and up to 50′ thick.	Enclosed in hornblende gneiss and lime silicate beds. Pegmatites cut foliation of host rock at high angle. Light gray augen gneiss at contact is silicified.	Pink microcline, rosy to gray quartz, graphic pegmatite, irregular bodies of dead white plagioclase. Beryl, small books of very black biotite, thin strips of biotite in feldspar cleavages.	Irregular texture along strike. Core some 2′ thick, mostly mined out, chiefly black quartz and beryl. Graphic pegmatite bordering core area. Pink microcline and albite, with quarts in border zones.	No outcrops of granitic rock in the area.	Has appearance, mineralogy and structure of a late Silver Plume dike.	Opened about 1942 and beryl cobbed out.
* Visited by M. F. Boos, July, 1939.							
#20. Hicks Mtn. beryl pegmatite, sec. 3, T. 5 S. R. 72 W. on E. side of summit, at about 9600′. Fig. 2.*	Irregular tabular dike 300′ long and 30′ thick. Strike nearly EW. Pods and lenses, branches and forks.	Enclosed in mica schist and knotty schist of Idaho Springs age. Dike follows foliation but branches cross the schistosity.	Smoky and glassy white quartz. Salmon-colored potash feldspar. Numerous small books of muscovite. Dull tan to white plagioclase. Small green beryl crystals and schorl.	Central part of exposed dike composed of white and smoky quartz, salmon-colored potash feldspar, many small books of muscovite. Dull tan plagioclase in wall zones. Beryl in quartz. Mica mostly in expanded parts of hanging wall.	Nearest outcrops of granitic rock are the Boulder Creek 1 mile to W. and Silver Plume to E. 1½ miles. No granite attached to dike.	Undetermined.	Not prospected.
* Visited by M. F. Boos, July 22, 1939.							
#21. Bald Mountain tourmaline pegmatite, sec. 9, T. 5 S., R. 72 W., opposite junction of Pedee and and Vance Creeks. Fig. 2.*	Lenticular dike with thick part at W. Trend nearly EW., 300′ long and up to 30′ thick. Lenses out to E. and disappears into slope. Vertical.	Enclosed in hornblende gneiss and schist, foliation of host N. 60° E., dips NW. low angle.	White "bull" quartz in a large central mass, pink to flesh-colored microcline, books of flawed muscovite. Light green prisms of tourmaline embedded in quartz. Schorl and an unknown bladed green mineral.	Dike pinches and swells out into schist so that mica foliae are flexed around pegmatite. Large core of white to black or gray quartz intergrown with blocky masses of microcline. Rod-like prisms of green tourmaline in quartz. Schorl and muscovite with albite surround the quarts core area.	Boulder Creek granite gneiss ½ mile to the W. Dike appears to be offshoot of Silver plume granite on Bald Mountain to the east.	Late Silver Plume age.	Not prospected.
* Examined by M. F. Boos, June 25, 1939.							
CLEAR CREEK CANYON AREA							
#22. Roscoe beryl pegmatite, sec. 5, T. 4 S., R. 71 W. Clear Creek Canyon. Fig. 2.*	Dike strikes N. 83° E., dips 75° SW., 100′ long 20′ thick. Vertical. Crosses Clear Creek.	Crosses foliation of injection gneiss and mica gneiss which here is EW.	Quartz, potash feldspar, biotite, and muscovite. Biotite in books up to 6″ and muscovite in books 2″ thick. Magnetite, garnet, small beryl prisms, gadolinite, monazite. Streaks of small red garnets.	Lobate margin, dike 35′ deep, widens below. Creamy microcline and graphic pegmatite around a core of quartz. Dull plagioclase in wall zones. Muscovite penetrates quartz in festoons of herring-bone books.	Sheets of Mount Morrison migmatite *lit-par-lit* with schist and gneiss of Idaho Springs beds.	Probably genetically related to the Mount Morrison gneiss.	Construction of new highway up Clear Creek canyon has destroyed a large part of the pegmatite.
* Visited by M. F. Boos, Jan., 1951.							
#23. Ajax mica pegmatite, sec. 34, T. 3 S., R. 72 W., on S.-facing slope of Clear Creek at about 7960′. Fig. 2.*	Irregular dike with several mica lodes. Body lies nearly EW., dips S. up to 40°. Outcrops for 300′. Walls roll on footwall side. Dike pinches out at SE. and is cut out by erosion at W.	Host rock mica schist foliated N. 50°–60° E. dips NW. steeply. Dike crosses foliation. Mica schist encloses the rolls along footwall. Schist dragged against pegmatite but pegmatite not folded. Books of muscovite lie in both pegmatite and host rock, transgressing both.	Microlite stains on feldspar. White and rosy quartz, albite, books and scrap type muscovite, also sunbursts of burr mica. Some greenish muscovite. Beryl, schorl, biotite. Muscovite books up to 14″ across and 4″ thick.	Well-defined zones with white to gray quartz core area, wall zones of light pink microcline thickly studded with small clear muscovite books. Border zone of 6″ on N. wall. Footwall a breccia of pegmatite and broken pieces of mica schist. Schist of hanging wall soaked in pegmatite, blended boundary. Schist altered and hardened for 2′ from pegmatite body.	Outcrop of Mount Morrison migmatite and injection gneiss on hillside about 200′ above and NE. of this pegmatite.	Developed within quarts-sillimanite mica schist and quarts-mica schist by the introduction of pegmatitic material, especially black quartz, schorl and mica. Partly reconstructed from the country rock.	Pegmatite mined for stove mica before 1900. About 1943 the mine was a source of scrap-mica for roofing. Both sheet and scrap-mica produced in 1944. Open cut 130′ by 40′–60′ and 40′ deep. At one time a drift on lower level.
* Examined by M. F. Boos and C. M. Boos, Dec. 24, 1945.							

† All locations referable to the 6th Principal Meridian.

tites. Pegmatitic material fills faults and sheared zones and partially occupies the contact zones of physically dissimilar rocks.

Magmatic invasion dominated the later granitic cycles. Silver Plume and Mount Olympus pegmatites fill finite fractures, occupy partings along the foliation and fissility, and penetrate the host rocks in incipient joint patterns. Magma, insinuated into fractured granite, developed intricate dike patterns. Some pegmatites make a halo around nongranitic inclusions that are buried in granite. The latest pegmatites fill joints, sheared zones and breaks. They resulted from the somewhat forceful injection of the fluid end-product of magma crystallization.

There is not much stress and little or no pegmatite where the granites of the earlier cycles adjoin. The emplacement of each pluton of Silver Plume granite created fractures that extend somewhat radially from the granite into host rocks. Adjustments of the area to the last of the Precambrian invasions (Boos and Aberdeen, 1940, p. 732) provided the loci for the late Mount Olympus pegmatites.

The shape and orientation of many pegmatites is related to the numerous minor structures in their host rocks. Low-dipping fractures in both granite and gneiss received fluids that created the well-zoned pegmatites. The western boundary of the Berrian Mountain pluton (Pl. 1) is streaked with pegmatites that fill late north-south fracture systems. Hematite stains and purplish color in quartz indicate water circulation in established breaks. Two types of pegmatites fill north-south fractures: those in granite knobs have low east dips and are a part of the body of the pluton and quartz-rich pegmatites that are vertical or dip west and occupy persistent fissures or fill tension breaks on the east flank of the Front Range.

There are many more pegmatites in the paragneisses and paraschists than in the granitic rocks. The lime-silicate and carbonate beds of the Idaho Springs harbor a good deal of pegmatitic material which occurs in the form of discontinuous pods and lenses (Pl. 5, fig. 3). Many of the dikes and sheets cross the marbles cleanly and do not seem to have greatly altered the adjacent calcareous beds.

Heinrich (1948, p. 442–448; 1953, p. 68–87) classifies the pegmatites that occur within granite bodies as interior; those in granite close to the contact with nongranitic rock as marginal, and those that are within rock outside of granite as exterior. The pegmatites of the Denver Mountain Parks area were developed within, marginal to, and adjacent to each stock, batholith, or pluton of the area. Probably a regional zonation pattern (the district zonation of Heinrich) of pegmatite development exists for each granite body with the exception of the smaller dikes and sills. However, no attempt is made in this paper to cite the mineralogical and structural features of the interior, marginal or exterior zones, partly because the close areal association of the different kinds of granite has produced an overprint of, for example, the exterior granites of one cycle on the exterior pegmatites of the earlier cycles.

Problems of Pegmatite Genesis

The thousands of granite pegmatites encountered during the field mapping of the Precambrian of the Denver Mountain Parks area provoked many questions concerning their occurrence, their distribution with respect to the granites, their genetic relationships to the five granitic rocks, and their place within host rocks. Some of the queries for which answers were sought in the field are:

(1) Do the pegmatites occur sporadically throughout the area, or are the pegmatites and associated aplites confined chiefly to the vicinity of the granitic bodies to which they are genetically related?

(2) Are the physical, mineralogical and structural features that distinguish the granites of one cycle from those of another reflected in their simple primary pegmatites?

(3) Have the pegmatites belonging to one cycle reliable, consistent intrinsic features that distinguish them from the pegmatites of earlier or later cycles?

(4) Does the structure and mineralogy of host rocks affect contained pegmatites?

(5) Do the pegmatites developed early in a granite cycle differ from those that developed late in the same cycle? If so, in what ways?

(6) Have the early pegmatites of the acid comagmatic region any distinctive features that are not native to pegmatites of the later granite cycles?

(7) Where do the scattered, complex, rare-

mineral-bearing pegmatites belong in the five granite cycles?

(8) Were the complex pegmatites the product of a single magmatic surge, or were they the result of repeated mineralizations induced by the successive granitic invasions of the area?

(9) Did some of the simple early primary pegmatites become "thoroughfares" or channelways for repeated replacements? Was the renewal of the granitizing and pegmatitizing processes a feature of succeeding granite cycles?

Primary Pegmatite Identities

Before the questions concerning the pegmatites could be answered, it was necessary to find means of relating the primary pegmatites and pegmatitic bodies to their respective parent bodies.

In areas of prolific pegmatite development, where more than one granite cycle had been operative, primary observations were made of one granite and its simple primary pegmatites. The following criteria were used to determine the characteristic intrinsic features that might prove the genetic relationships of the pegmatites and their parent granites.

(1) Interconnection of pegmatite and granite on the outcrops or in very close areal association.

(2) Physical and structural features common to granite and associated pegmatites.

(3) Characteristic minerals present in both.

(4) Relative amounts of essential minerals.

(5) Color, crystal patterns and associations of the accessory minerals.

(6) Occurrence of primary zonation and association of quartz, mica, and feldspars in zones.

(7) Sequence of emplacement.

The distinctive characteristics of pegmatites enclosed by granite were compared with pegmatites having nongranitic hosts. Where the evidence was available, internal structures of pegmatites, such as zonation and alignment of mineral units were compared for pegmatites of the different granite cycles.

Characteristic Features of Primary Simple Pegmatites and the Granites of each Cycle

Mount Morrison cycle.—Primary pegmatites and the Mount Morrison gneiss have many of the following features:

(1) The pegmatites range from micro-sheets and pencils enclosed within the Mount Morrison migmatite and gneiss to tabular bodies hundreds of feet long and tens of feet thick that finger out from injection gneiss and migmatite. Mixed gneiss-aplite-pegmatite sheets border the main Mount Morrison migmatite outcrops. Densely feldspathic pegmatites occur as termini of many sheets of Mount Morrison gneiss.

(2) The pegmatites, like the Mount Morrison gneiss, are generally gneissic (Pl. 2, fig. 3). Discontinuous clusters of similarly-oriented, intergrown biotite and muscovite flakes make rather evenly-spaced layers that roughly parallel the walls of the pegmatite or follow the trend of the foliation of the enclosing rocks. The small pegmatites are more obviously gneissic than the large ones. The earliest pegmatites of the Mount Morrison cycle are distinctly gneissic, but later ones are less so.

(3) The chief minerals of the Mount Morrison gneiss and its pegmatites are primary albitic plagioclase, microcline, and orthoclase. The crystals are anhedral and are much granulated. Graphic textures are rare. The pegmatites are quartz-poor. Quartz is more plentiful in the Deer Creek area than elsewhere and is a peculiar bluish gray. Quartz in the pegmatites of the Mount Vernon area is greenish, and some is purple-stained.

(4) The fresh pegmatites are creamy to white. The light color of both pegmatite and Mount Morrison gneiss is due to the large amount of plagioclase feldspar and some light-colored potash feldspar. When weathered, the feldspars of the Mount Morrison pegmatites become dull and cleave into a *gruss* of crumbled particles.

(5) White micas and book biotite are generally absent in both gneiss and pegmatite. There is little or no schorl in the pegmatites, but their nongranitic host rocks are black with felted layers of crowded schorl needles. Grains and lumps of magnetic iron oxide, intergrown with brown garnets and orthite, are persistent and widespread accessories of both the Mount Morrison gneiss and its pegma-

tites. Octahedra of magnetite are frozen in the quartz, especially where the host rock is hornblende gneiss.

(6) The early Mount Morrison pegmatites are not zoned. A few late Mount Morrison pegmatites have a crude simple zonation (Pl. 2, fig. 4). The zoned pegmatites are not gneissic. Quartz is the essential mineral of the cores. Wall zones are composed of massive feldspar. Inclusions of schist, elongated parallel to the foliation of host rocks, occur sparingly.

(7) The Mount Morrison pegmatites are crossed by Boulder Creek, Pikes Peak and Silver Plume pegmatites.

Pegmatites of Mount Morrison age are exposed on Mount Falcon, over Shaffer Hill, in irregular masses that grade into Mount Morrison gneiss on Saddleback Mountain, and north of Cody Park (Pl. 1).

At least three generations of Mount Morrison pegmatites outcrop on the south side of Mount Morrison within the gneiss and migmatite, and adjacent to it in the wide zone of injection gneiss that composes the southern part of the mountain. The earliest Mount Morrison pegmatites are feldspathic and gneissic. They are warped and folded with the Idaho Springs beds around them. They consist chiefly of albitic feldspar and are poor in quartz. The large dikes contain haphazardly-scattered pods of yellow quartz.

Pegmatites contemporaneous with the Mount Morrison development on Mount Morrison blend with it and are simply coarse-grained variants of the gneiss. They consist of ill-defined lenses of red-stained feldspar and some quartz. Clusters of intergrown orthite, biotite, and brown garnet are the chief accessory minerals. The latest pegmatites of the area extend from the coarser, delicately-banded gneiss into the injection gneiss, crossing the gneissosity at a high angle. They are quartz-rich, selvaged by potash feldspar, and mostly trend northeast of north. They cross the narrow aplite dikes that preceded them and branch out into quartz veins. Where the late dikes are enclosed in mica schist, the rock at the contact is thickly impregnated with needles of schorl. In some places, the late dikes have a recognizable quartz core that has in it a little white mica and minute books of biotite.

A few pegmatites that originated in Mount Morrison times have been mined for feldspar and for quartz. The primary pegmatite has a low mica content. A few dikes among the many hundreds belonging to the cycle contain beryl, tantalum ore, and other rare minerals (Table 1, nos. 2, 4, 9, 17, 23), but their complex character and mineralization is probably of later than Mount Morrison times.

Boulder Creek cycle.—Primary pegmatites and the Boulder creek granite gneiss of the Boulder Creek cycle have many of the following features:

(1) Most of the pegmatites fill sheared fractures and joints in the Mount Morrison gneiss and the Boulder Creek gneiss. Some also occupy partings in the Idaho Springs beds, but *lit-par-lit* intrusion is rare. Boulder Creek pegmatites occur as schlieren and as coarse, lensing segregations all through the granite gneiss. The pegmatites have associated stringers and sheets of aplite.

(2) The pegmatites are not primarily gneissic. Where they traverse the metasedimentary rocks, they contain streaks of mica schist. Any banded or streaky appearance is not due to preferred orientation of primary pegmatite minerals. Some dikes are massive; others are zoned.

(3) The pegmatites, like the parent Boulder Creek granite and granite gneiss, contain two generations of biotite.

(4) The chief minerals of both Boulder Creek granite and its pegmatites are oligoclase, microcline, orthoclase, white to gray quartz, and biotite. The all-over color is tan to pale pink owing to the large content of light-colored plagioclase and flesh-colored potash feldspar. Boulder Creek pegmatites contain more quartz than pegmatites of Mount Morrison origin. Muscovite is rare.

(5) Accessory minerals of both the Boulder Creek and its pegmatites are hornblende, dusty-brown garnet, orthite, hematite, and magnetite. Primary beryl is rare. The minor minerals occur all through the pegmatites and in about the same proportions, relative to essential minerals as the accessories of the Boulder Creek granite and gneiss bear to the chief con-

mineral-bearing pegmatites belong in the five granite cycles?

(8) Were the complex pegmatites the product of a single magmatic surge, or were they the result of repeated mineralizations induced by the successive granitic invasions of the area?

(9) Did some of the simple early primary pegmatites become "thoroughfares" or channelways for repeated replacements? Was the renewal of the granitizing and pegmatitizing processes a feature of succeeding granite cycles?

Primary Pegmatite Identities

Before the questions concerning the pegmatites could be answered, it was necessary to find means of relating the primary pegmatites and pegmatitic bodies to their respective parent bodies.

In areas of prolific pegmatite development, where more than one granite cycle had been operative, primary observations were made of one granite and its simple primary pegmatites. The following criteria were used to determine the characteristic intrinsic features that might prove the genetic relationships of the pegmatites and their parent granites.

(1) Interconnection of pegmatite and granite on the outcrops or in very close areal association.

(2) Physical and structural features common to granite and associated pegmatites.

(3) Characteristic minerals present in both.

(4) Relative amounts of essential minerals.

(5) Color, crystal patterns and associations of the accessory minerals.

(6) Occurrence of primary zonation and association of quartz, mica, and feldspars in zones.

(7) Sequence of emplacement.

The distinctive characteristics of pegmatites enclosed by granite were compared with pegmatites having nongranitic hosts. Where the evidence was available, internal structures of pegmatites, such as zonation and alignment of mineral units were compared for pegmatites of the different granite cycles.

Characteristic Features of Primary Simple Pegmatites and the Granites of each Cycle

Mount Morrison cycle.—Primary pegmatites and the Mount Morrison gneiss have many of the following features:

(1) The pegmatites range from micro-sheets and pencils enclosed within the Mount Morrison migmatite and gneiss to tabular bodies hundreds of feet long and tens of feet thick that finger out from injection gneiss and migmatite. Mixed gneiss-aplite-pegmatite sheets border the main Mount Morrison migmatite outcrops. Densely feldspathic pegmatites occur as termini of many sheets of Mount Morrison gneiss.

(2) The pegmatites, like the Mount Morrison gneiss, are generally gneissic (Pl. 2, fig. 3). Discontinuous clusters of similarly-oriented, intergrown biotite and muscovite flakes make rather evenly-spaced layers that roughly parallel the walls of the pegmatite or follow the trend of the foliation of the enclosing rocks. The small pegmatites are more obviously gneissic than the large ones. The earliest pegmatites of the Mount Morrison cycle are distinctly gneissic, but later ones are less so.

(3) The chief minerals of the Mount Morrison gneiss and its pegmatites are primary albitic plagioclase, microcline, and orthoclase. The crystals are anhedral and are much granulated. Graphic textures are rare. The pegmatites are quartz-poor. Quartz is more plentiful in the Deer Creek area than elsewhere and is a peculiar bluish gray. Quartz in the pegmatites of the Mount Vernon area is greenish, and some is purple-stained.

(4) The fresh pegmatites are creamy to white. The light color of both pegmatite and Mount Morrison gneiss is due to the large amount of plagioclase feldspar and some light-colored potash feldspar. When weathered, the feldspars of the Mount Morrison pegmatites become dull and cleave into a *gruss* of crumbled particles.

(5) White micas and book biotite are generally absent in both gneiss and pegmatite. There is little or no schorl in the pegmatites, but their nongranitic host rocks are black with felted layers of crowded schorl needles. Grains and lumps of magnetic iron oxide, intergrown with brown garnets and orthite, are persistent and widespread accessories of both the Mount Morrison gneiss and its pegma-

tites. Octahedra of magnetite are frozen in the quartz, especially where the host rock is hornblende gneiss.

(6) The early Mount Morrison pegmatites are not zoned. A few late Mount Morrison pegmatites have a crude simple zonation (Pl. 2, fig. 4). The zoned pegmatites are not gneissic. Quartz is the essential mineral of the cores. Wall zones are composed of massive feldspar. Inclusions of schist, elongated parallel to the foliation of host rocks, occur sparingly.

(7) The Mount Morrison pegmatites are crossed by Boulder Creek, Pikes Peak and Silver Plume pegmatites.

Pegmatites of Mount Morrison age are exposed on Mount Falcon, over Shaffer Hill, in irregular masses that grade into Mount Morrison gneiss on Saddleback Mountain, and north of Cody Park (Pl. 1).

At least three generations of Mount Morrison pegmatites outcrop on the south side of Mount Morrison within the gneiss and migmatite, and adjacent to it in the wide zone of injection gneiss that composes the southern part of the mountain. The earliest Mount Morrison pegmatites are feldspathic and gneissic. They are warped and folded with the Idaho Springs beds around them. They consist chiefly of albitic feldspar and are poor in quartz. The large dikes contain haphazardly-scattered pods of yellow quartz.

Pegmatites contemporaneous with the Mount Morrison development on Mount Morrison blend with it and are simply coarse-grained variants of the gneiss. They consist of ill-defined lenses of red-stained feldspar and some quartz. Clusters of intergrown orthite, biotite, and brown garnet are the chief accessory minerals. The latest pegmatites of the area extend from the coarser, delicately-banded gneiss into the injection gneiss, crossing the gneissosity at a high angle. They are quartz-rich, selvaged by potash feldspar, and mostly trend northeast of north. They cross the narrow aplite dikes that preceded them and branch out into quartz veins. Where the late dikes are enclosed in mica schist, the rock at the contact is thickly impregnated with needles of schorl. In some places, the late dikes have a recognizable quartz core that has in it a little white mica and minute books of biotite.

A few pegmatites that originated in Mount Morrison times have been mined for feldspar and for quartz. The primary pegmatite has a low mica content. A few dikes among the many hundreds belonging to the cycle contain beryl, tantalum ore, and other rare minerals (Table 1, nos. 2, 4, 9, 17, 23), but their complex character and mineralization is probably of later than Mount Morrison times.

Boulder Creek cycle.—Primary pegmatites and the Boulder creek granite gneiss of the Boulder Creek cycle have many of the following features:

(1) Most of the pegmatites fill sheared fractures and joints in the Mount Morrison gneiss and the Boulder Creek gneiss. Some also occupy partings in the Idaho Springs beds, but *lit-par-lit* intrusion is rare. Boulder Creek pegmatites occur as schlieren and as coarse, lensing segregations all through the granite gneiss. The pegmatites have associated stringers and sheets of aplite.

(2) The pegmatites are not primarily gneissic. Where they traverse the metasedimentary rocks, they contain streaks of mica schist. Any banded or streaky appearance is not due to preferred orientation of primary pegmatite minerals. Some dikes are massive; others are zoned.

(3) The pegmatites, like the parent Boulder Creek granite and granite gneiss, contain two generations of biotite.

(4) The chief minerals of both Boulder Creek granite and its pegmatites are oligoclase, microcline, orthoclase, white to gray quartz, and biotite. The all-over color is tan to pale pink owing to the large content of light-colored plagioclase and flesh-colored potash feldspar. Boulder Creek pegmatites contain more quartz than pegmatites of Mount Morrison origin. Muscovite is rare.

(5) Accessory minerals of both the Boulder Creek and its pegmatites are hornblende, dusty-brown garnet, orthite, hematite, and magnetite. Primary beryl is rare. The minor minerals occur all through the pegmatites and in about the same proportions, relative to essential minerals as the accessories of the Boulder Creek granite and gneiss bear to the chief con-

stituents (Boos, 1935, p. 1037). Magnetic iron-oxide masses are rare: schorl has not been noted.

(6) The dikes have veinlike cores of glassy-white quartz and some potash feldspar. Pegmatites that are of late Boulder Creek age are coarsely zoned (Pl. 3, fig. 2). Graphic textures are common. Most dikes have sharp boundaries against host rocks.

(7) Boulder Creek pegmatites cut pegmatites and gneiss of Mount Morrison age.

Pegmatites of early Boulder Creek genesis are massive, white, feldspathic, and biotite-rich. Pegmatites contemporaneous with the Boulder Creek granite crystallization are simple, giant-grained Boulder Creek bodies. The pegmatites of late Boulder Creek genesis have veinlike quartz cores. Wall zones are chiefly feldspar. The margins carry narrow, fine-grained border zones. Small books of bright black biotite occur throughout the pegmatite (Pl. 5, fig. 3).

Boulder Creek pegmatites and aplites outcrop along the Mount Evans highway above Echo Lake and within the Boulder Creek granite and granite gneiss. The dikes along the Squaw Pass road, near Squaw Pass, have biotite-sillimanite schist and knotenschiefer as hosts. The pegmatites genetically related to the Medlen School and Fairy Hill stocks of Boulder Creek have been prospected for feldspar (Table 1, nos. 7, 14, 15, 16, 18).

Pikes Peak cycle.—Primary pegmatites and the Pikes Peak granite of the Pikes Peak cycle have the following features:

(1) The pegmatites occur as schlieren and as discrete dikes within the Pikes Peak granite or adjacent to it. A few dikes that appear to be of Pikes Peak origin outcrop where no Pikes Peak granite is at the surface. No Pikes Peak pegmatites outcrop north of Bear Creek.

(2) The spotted appearance of the Pikes Peak granite (Pl. 3, fig. 4) is duplicated, faintly, in some of the pegmatites and aplites of Pikes Peak genesis. Clots of biotite and hornblende are scattered irregularly in the dikes. The feldspars are anhedral and somewhat granulated. The pegmatites and the Pikes Peak granite are not gneissic.

(3) and (4) The chief minerals of both Pikes Peak granite and its pegmatites are pink to red microcline, white to gray plagioclase, white quartz, and one or more generations of biotite or hornblende. Primary muscovite is not common.

(5) The accessory minerals that characterize the Pikes Peak granite appear in the pegmatite also, but in smaller amounts relative to the chief minerals. Schorl occurs sparingly. Purple fluorite, the most reliable diagnostic accessory mineral of the Pikes Peak pegmatite is present in almost all of the dikes as it is in the Pikes Peak granite (Boos, 1935, p. 1039). Sillimanite is a late mineral. The bright red to salmon color of the pegmatite and the Pikes Peak granite in many places is due to the numerous inclusions of blood-red hematite in the potash feldspars. The red color is striking, adjacent to dead white quartz. Gray feldspar in the Pikes Peak pegmatite is usually associated with bluish to milky white quartz.

(6) Most of the pegmatites of Pikes Peak genesis are zoned. Typical dikes carry cores of dead-white quartz intergrown with salmon-colored potash feldspar. The wall zones are composed of creamy plagioclase, books of rusty biotite, and minor amounts of red potash feldspar. Black schorl is frozen in white quartz. Masses of graphic microcline are plentiful.

(7) Pegmatites that contain characteristic Pikes Peak features invade pegmatites of Mount Morrison age.

The pegmatites that developed late in the Pikes Peak cycle have well-defined zones. The cores consist of red feldspar and white quartz flanked by wall zones with graphic textures. In places, intermediate zones contain quartz, a plagioclase, fluorite, and nests of schorl. Fine-grained border zones are persistent. The Pikes Peak pegmatites that are contemporaneous with the granite are chiefly pods, lenses and coarse-grained stringers that blend with the normal granite.

Pegmatites that contain some of the features characteristic of Pikes Peak genesis occupy fractures that strike N. 70° W. in gneiss near the old fluorite mine south of Evergreen. Both

purple and green fluorite occur in the pegmatites, and all the veins of fluorite are pegmatitic. Pegmatites welded to Pikes Peak granite, on the east slope of Black Mountain, are chiefly aggregates of red feldspar and dead-white quartz. Pegmatites of Pikes Peak aspect occupy the contact of Boulder Creek and Silver Plume granites and finger out into the Boulder Creek gneiss in the Indian Creek Park area. Nonzoned, spotted pegmatites of early Pikes Peak genesis outcrop near the highway at the entrance to Harris Park south of the map area of Pl. 1.

Silver Plume-Mount Olympus cycle.—Primary pegmatites and the Silver Plume and Mount Olympus granites of the Silver Plume-Mount Olympus cycle have many of the following features:

(1) The pegmatites and aplites occur within Silver Plume granite or are associated with outcrops of Silver Plume or Mount Olympus granite. They inject the formations bordering the Indian Creek plutons and constitute a large part of the injection gneiss that encloses them. Most of the pegmatites appear to have been introduced magmatically.

(2) The pegmatite, like the Silver Plume and Mount Olympus granites, are neither gneissic nor are they streaked with inclusions. Those enclosed in granite are less distinctly zoned than those that have host rock of schist or gneiss.

(3) The mineral grains of both pegmatites and Silver Plume granite are fresh, distinct, and relatively large. Few grains are appreciably strained or granulated. In most dikes, grain size increases progressively from walls to the center of the pegmatite. Where Silver Plume and Mount Morrison dikes outcrop in the same area, the unweathered appearance of pegmatites of Silver Plume genesis contrasts sharply with the crumbly, dull, broken appearance of the light-colored pegmatites of Mount Morrison age which superficially resemble them. Pegmatites of Silver Plume-Mount Olympus genesis retain their fresh appearance and shape on the outcrop.

(4) The chief minerals of the pegmatites and parent granites are fresh-colored microcline of pearly luster, gray to almost black quartz, and small amounts of creamy to tan plagioclase. Both granite and pegmatite have a clean fresh appearance due to the large content of unaltered potash feldspar. Books of silvery-white mica, mostly flawed muscovite, are associated with quartz in the pegmatites. Festoons of grayish to silvery muscovite with herringbone pattern fill fractures in quartz and cleavages in the feldspars.

(5) The accessory minerals consist of red garnet, small books of biotite, octahedra of magnetite, schorl and minute hyacinth zircons. Orthite and apatite are very rare (Boos, 1935, p. 1042). In general, the pegmatites, like the Silver Plume and Mount Olympus granites, have fewer accessory minerals and less of them than the pegmatites of older cycles.

(6) Most Silver Plume and Mount Olympus pegmatites are coarsely holocrystalline and zoned. They contain cores of quartz and potash feldspar, wall zones of microcline, graphic microcline and smoky quartz and some muscovite. The fine-grained border zones consist of quartz, plagioclase and a little white mica. Clusters of small radiating schorl prisms extend from the border zones inward.

(7) Silver Plume pegmatites cut across Boulder Creek and Pikes Peak pegmatite and granite bodies.

Aplite and pegmatite are interlayered in some dikes. Next to the Mount Morrison aplite, that genetically related to the Mount Olympus granite is the most abundant of all late Precambrian aplites.

Pegmatites of Silver Plume and of Mount Olympus genesis are strikingly similar. The chief distinction between the two granites is in their texture and mineral orientations. This difference is not reflected in their respective pegmatites.

Many of the pegmatites are pod-shaped or lenticular. Funnelform masses in granite widen upwards to a massive quartz "cap." Others enlarge, on the strike, to a bluntly-terminated head. Pegmatites make an irregular halo around inclusions of schist and gneiss within granite. None of the pegmatites is ptygmatically folded.

The pegmatites are, for the most part, tabular bodies that have little or no distortion. Only those in the mica schists pinch and swell. Silver Plume and Mount Olympus pegmatites enclosed in mica schist contain much more muscovite and less plagioclase than those embedded in granite or quartz gneiss. Biotite and magnetite are usually present where hornblende gneiss is the host.

North and northeast-trending fissures (Pl. 1, Secs. 4, 5, 8, 9, of T. 5 S., T. 70 W.) that extend from the Boulder Creek plutons are healed, locally, with Silver Plume pegmatite [Pl. 1]. Some of these dikes, in the Parmalee Gulch area, have been sheared by later movements along the original fracture systems.

Silver Plume pegmatites appear along the highway between Indian Hills and Conifer on the east side of the Doublehead pluton. A network of them outcrops in the gneiss bordering the Troutdale pluton (Sec. 4, T. 5 S., R. 71 W). The schist and gneiss adjacent to the east side of the Hodgson School pluton are bordered by long, thick subparallel dikes and sills, especially in secs. 20, 21, 28, T. 5 S., R. 70 W. Mount Olympus pegmatites and granites outcrop on Sampson Mountain and Huebner Peak and along the highway west of Critchell. Many well-zoned pegmatites of probable Mount Olympus age outcrop in the hills north of Kittredge.

Pegmatites genetically related to the Silver Plume and Mount Olympus granites are zoned. There is little to distinguish the latter ones from those that immediately preceded the introduction of the Silver Plume granite. The intermediate pegmatites of the cycle are granitic and coarsely crystallized. They occupy partings and fractures in both granites and metamorphics.

The latest primary pegmatites of the Mount Olympus part of the cycle are tabular bodies that have well-defined, gray to black quartz-rich cores, wall zones with abundant, pearly, flesh-colored microcline, silvery muscovite books that are badly flawed, no biotite, and narrow "chill" borders containing fine-grained quartz, plagioclase, and minute flakes of muscovite intergrown with minute prisms of schorl. Rosettes of schorl prisms occur in the lighter-colored quartz of the core area.

Within a pegmatite-rich area of Silver Plume genesis dikes range from almost pure quartz vein pegmatite to quartz-rich dikes with selvage zones of potash feldspar, to well-zoned pegmatites consisting of a quartz core, feldspar wall zones with units of graphic texture and fine-grained border zones, to dikes in which the quartz core makes up less than 10 percent of the pegmatite. The last to be developed during the fifth cycle of granitic rocks consist of white vein quartz sheets that trend northeast and dip northwest.

Quartz Diorite

The small exposures of quartz diorite in Turkey Creek and Deer Creek canyons have little or no pegmatitic material associated with them. Practically nothing is known of the pegmatitic material genetically related to this diorite in the Denver Mountain Parks area.

Summary

Typical simple primary pegmatites of Mount Morrison age are gneissic, nonzoned, contain wispy relicts of host rocks, lack nonferrous accessory minerals, and usually carry lumps of magnetite. They are dull, light-colored, rich in primary plagioclase and lesser amounts of potash feldspar, lack micas, crumble readily on exposure, and are generally poor in quartz. Magnetite is the abundant and persistent accessory mineral.

The pegmatites range from closely-spaced, parallel pencils in the migmatitic phases of the Mount Morrison gneiss to long massive sheets in injection gneiss. The smaller dikes are ptygmatically folded. The others are distorted with the host rocks. The pegmatites, with associated aplites, were the great *lit-par-lit* invaders of the Idaho Springs beds in the Bear creek and Clear Creek drainage areas.

Typical simple primary pegmatites of Boulder Creek age are not gneissic. Many of them are massive and undifferentiated but others are marked by vein-like quartz cores. Streakiness is due to relicts of host rocks. The pegmatites are chiefly tabular bodies with sharp contacts against hosts. There are nearly equal amounts of orthoclase and plagioclase. White to glassy gray quartz is abundant locally. Biotite is always present, usually in two generations of books and sheets. The earlier biotite

consists of fresh black books, and the later biotite occurs in strips and thin sheets that fill the contacts of mineral units. Accessory magnetite intergrown with brown rusty garnets and orthite is common.

Typical simple primary pegmatities of Pikes Peak genesis are zoned. Some of them are crudely layered and contain quartz-feldspar cores. Neither gneissic structures nor country-rock inclusions are present. The potash feldspar is usually salmon red, and the lime-soda feldspars are gray. All the feldspar is usually fresh and colorful. Quartz is dead white to milky. The appearance of an imperfectly-zoned dike is spotted due to sharp color variations within the feldspars and to scattered patches of hornblende and biotite. Fluorite is a diagnostic accessory mineral. Orthite, sphene, biotite, muscovite, and apatite have been found.

Typical simple primary pegmatites of Silver Plume-Mount Olympus genesis occur in tabular bodies or pinch-and-swell pods. Few of them show folding or bending. All are distinctly and consistently zoned. They are present wherever Silver Plume or Mount Olympus granite occur. Many of them intrude the metamorphic rocks that border the Indian Creek plutons. A few extend somewhat radially out from the granite into host rocks from the plutons. They fill fissures that are steeply inclined. They are bright, fresh-appearing and lustrous, due to their large content of unweathered microcline. Gray to black quartz is abundant.

Diagnostic accessory minerals include schorl in well-defined prisms, red garnets, hyacinth zircon, and small octahedra of magnetic iron. Muscovite and a little fresh biotite are associated with quartz. Muscovite characterizes the dikes enclosed in mica schist.

Silvery muscovite occurs in three of four pegmatites of Silver Plume-Mount Olympus genesis. The muscovite of Boulder Creek pegmatites is usually grayish. The very rare white mica of the Mount Morrison pegmatites lacks a silvery luster.

There is a greater similarity among the pegmatites that have a host rock of granite than between those enclosed in schist or gneiss. In general, those within granite are giant-grained granites or simple-zoned dikes that grade into granite. Those that transgress the foliation of schistosity of metasediments are more obviously zoned that those that conform to the bedding or foliation of the rocks enclosing them.

Pegmatites enclosed in mica schist usually contain fronds and festoons of flawed muscovite books or ball-like masses of shot-size muscovite. The festoons dip into the wall zones from the inner edge of the border zones. Minute flakes of white mica pepper the borders.

Complex Pegmatites

The foregoing pages deal chiefly with the simple, primary Precambrian granitic pegmatites of the Denver Mountain Parks region. In some of the pegmatite-rich areas (Fig. 2) a few complex pegmatites (Landes, 1933) are associated with the common simple ones. A small number of the complex pegmatites contain potentially valuable commercial bodies of quartz, feldspar, and mica. Some carry beryl, radioactive compounds, tantalum ores, and in at least one case there is bismuth ore (Table 1). No two of the complex pegmatites, however, are alike in composition or in their relationships to parent granites.

Several areas within the granite-rich parts of the mapped portion (Pl. 1) contain extensive swarms of both simple and complex pegmatites some of which have been mined or prospected. Many of the larger unexplored bodies outcrop near communities of closely-spaced summer homes where exploitation of the ores would be unfeasible at this time.

Some 75 of the hundreds of large pegmatites have been opened by mining or prospecting so that a three-dimensional picture has been obtained by year-to-year mapping of the pegmatite body exposed. Information concerning the content, internal structure, relationships to host rocks, and probable source has been obtained for several pegmatites, details of which are given in Table 1.

Geologic Associations of the Complex Pegmatites

Nineteen complex pegmatites (Table 1) are either closely associated with outcrops of Mount Morrison formation or stem directly from it. Among them are the Snyder, Grover, Coors, Ajax, and east Shaffer Hill dikes (Figs. 3, 5, 6). Both Mount Morrison and Boulder Creek either outcrop near or are in contact with

FIGURE 3.—MAP OF GROVER PEGMATITE AND MINE

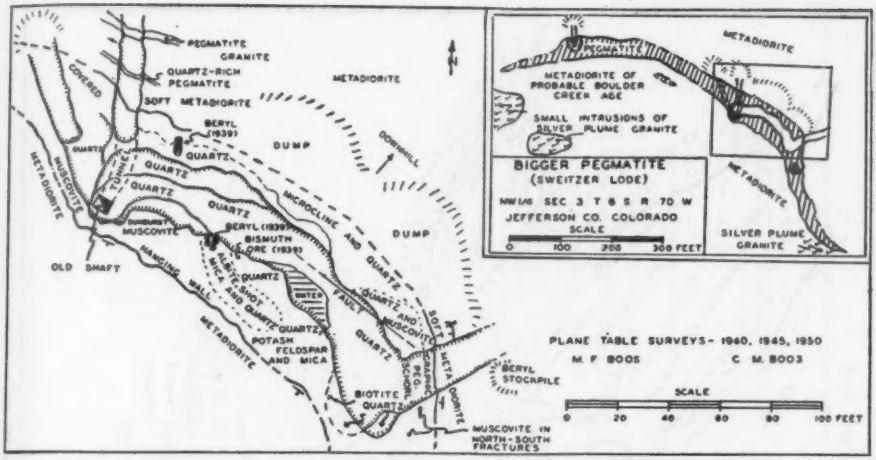


FIGURE 4.—MAPS OF BIGGER PEGMATITE AND MINE

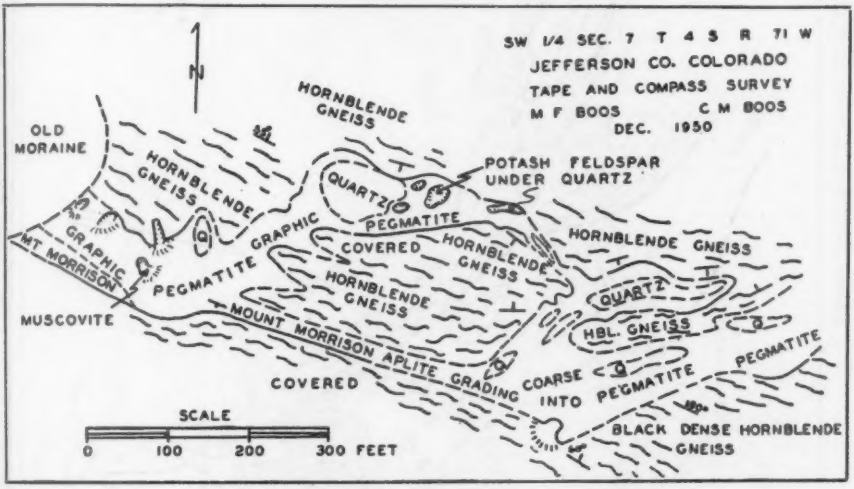


FIGURE 5.—MAP OF PEGMATITES ON EAST SHAFFER HILL

the Beaver Brook School, Old Soda Creek School, Thomas and Fenders pegmatites. The Medlen School and Bigger pegmatites are associated with both Boulder Creek gneiss and Silver Plume granite (Fig. 4).

Sunrise Peak, Silver Glen Ranch, Swede Gulch, and Wasson pegmatites are closely associated with both the Mount Morrison and the Silver Plume granites (Figs. 7, 8). The Mount Morrison is actually a part of the body of some pegmatites (Figs. 6, 7, 8) and the Silver Plume intruded the dike itself or outcrops near by. The relatively simple pegmatites known as Bald Mountain, Iowa Gulch and Seven Hills have only the Silver Plume granite as a "mother rock" (Table 1, nos. 11, 17, 21).

Twenty-one of the pegmatites in Table 1 have well-developed zonation. Twenty have steep

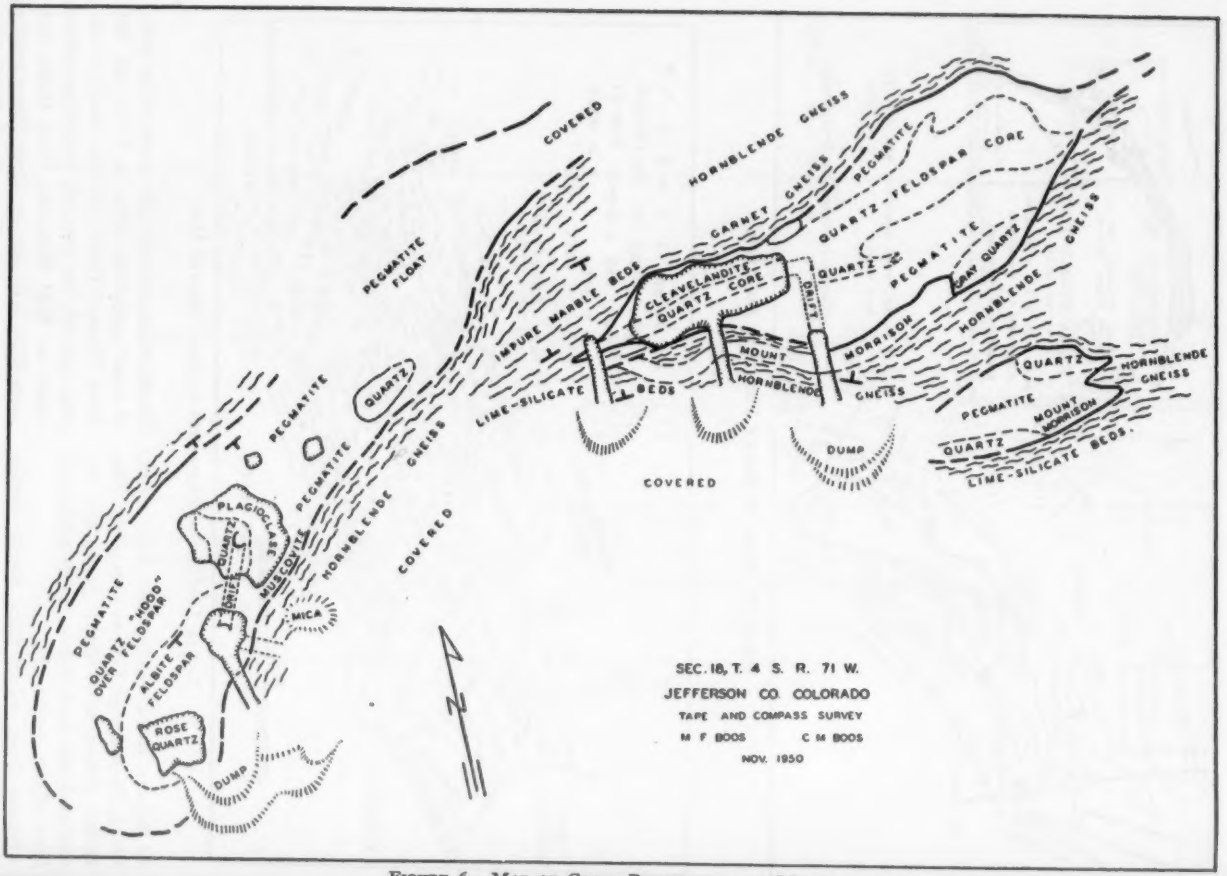


FIGURE 6.—MAP OF COORS PEGMATITES AND MINES

FIGURE 6.—MAP OF COORS PEGMATITES AND MINES

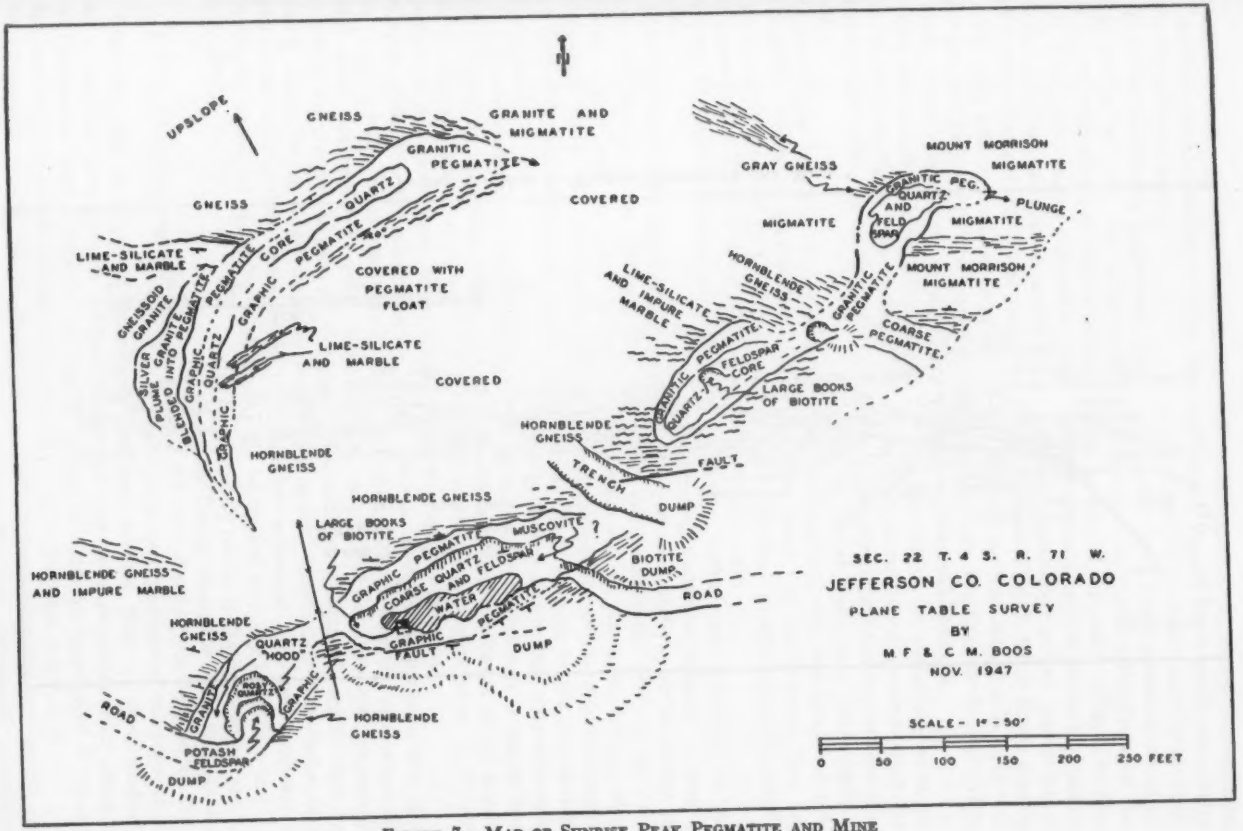


FIGURE 7.—MAP OF SUNRISE PEAK PEGMATITE AND MINE

FIGURE 8.—MAP OF WASSON PEGMATITE AND MINE

dips and cross the foliation of host rocks locally. Few conform to the foliation or follow the fissility of their host rocks.

Almost all of the complex pegmatites contain colored quartz, but very few of the simple primary ones carry other than white or colorless quartz. Ten of the complex pegmatities given in Table 1 have rose quartz cores, and 15 have black to gray quartz chiefly in the core zone. In a few dikes some of the black quartz is of later origin than the gray or white. The greenish or bluish quartz characteristic of the Mount Morrison granitic rocks occurs in a few complex dikes. Colored quartz is best developed in pegmatite dikes that are exceptionally well zoned.

Colored quartz is usually accompanied by cleavelandite. Albite or cleavelandite of late origin is present in more than half of the complex pegmatites of Table 1. Most pegmatites that carry beryl also contain secondary plagioclase.

White beryl, with white cleavelandite, occurs in the intermediate zones. Thirteen dikes of Table 1 contain beryl in either the wall or intermediate zones. All pegmatites, closely associated with the Boulder Creek gneiss, carry some beryl.

It seems reasonable that a complex pegmatite, developed in its present entirety during, say, the Mount Morrison cycle, would closely resemble mineralogically any other complex pegmatite of the same generation. However, no two of the complex pegmatites observed are either alike structurally or carry identical mineral units of rare mineral groups. The simple granitic pegmatites of a granite cycle have features that indicate their consanguinity, but the complex ones show no family resemblance in the field.

Genesis of Certain Complex Pegmatites

Generation after generation of acid magma surges, occurring in rather close succession in late Precambrian times, developed two batholiths, some stocks, many plutons, and dikes and sills composed of granitic rock in the Denver Mountain Parks area.

Field mapping of the granitic and metamorphic rocks (Spurr, Garrey and Ball, 1908, Pl. 11; Pl. 1 of this paper) shows that the Mount Evans batholith of Boulder Creek granite and granite gneiss is intruded by small bodies of Pikes Peak granite and Silver Plume granite. The pegmatites of the later granite cycles also cross the granitic rocks of the earlier cycles. The very close areal association of all the kinds of granite and their genetically related pegmatites is plainly shown on Pl. 1. Silver Plume granite, for example, has been identified in dikes that cross the Mount Morrison, the Boulder Creek, and the Pikes Peak outcrops.

The introduction of pegmatites, wave on wave, went on with quiet periods between. At great depths the pegmatitization processes were probably almost continuous. The originally deep-seated granite areas now compose a considerable portion of the outcrops that are riddled with simple granitic pegmatites, chiefly of magmatic origin.

However, action that originated in the country rock under the influence of magma resulted in granitization (metasomatism as defined by Goldschmidt, 1922, p. 105–123), ultrametamorphism (Hietanen, 1951, p. 605), metamorphic differentiation, and general palingenesis. Hence, changes in the granitic rock already in place and in the nongranitic rocks of the area under the influence of magma heat, pressure, and fluids resulted in aplites, pegmatites, and those pegmatitic veins not directly caused by magma intrusion.

Jahns noted for the pegmatites of New Mexico (1946, p. 73) "as more and more pegmatites are investigated it becomes increasingly apparent that no single theory of origin can have general application." The author is in complete agreement with this statement. It can be applied to many pegmatites that are genetically related to the Precambrian granites of the Denver Mountain Parks area but that do not stem directly from the granites. However, it is not the purpose of this paper to distinguish carefully those of direct magmatic origin from those of metamorphic origin, although all kinds are represented in the area (Pl. 3, fig. 2; Pl. 2, figs. 2, 4, 3, 5, 6).

Each magma responsible for a granite cycle seems to have had characteristic constituents that produced like accessory minerals in the granite and its pegmatites. This does not rule

out the possibility that minerals crystallizing from the residual liquors may occur in some of the pegmatites of a cycle and not in others. Mineralization of most pegmatites was completed within the single granite cycle to which they belonged.

Plagioclase—usually cleavelandite—colored quartz, sheet and burr mica, and logs of beryl in a complex pegmatite could be either the product of the magmatic activity of a single granitic cycle, or might have been introduced in a subsequent acid magma period. Rare earth minerals seem to be the latest additions to a complex pegmatite. The writer believes that the additions to primary pegmatite and the alterations and replacements were due to more than one cycle of acid magma action. Thus the Bigger pegmatite (Table 1, no. 14), initially of Boulder Creek genesis, has probably been enriched by the magma of a later granite cycle. The large, early beryl prisms were corroded, bismuth minerals were introduced, and a late generation of silvery muscovite, characteristic of the Silver Plume cycle, developed within the body of the original simple pegmatite (Pl. 5, figs. 4, 5).

Opportunities for the replacements of primary granitic pegmatites by minerals of a later cycle were at least four-fold. Any one of the later Precambrian granitic invasions of the area might have enforced changes in the pegmatites of an earlier cycle. Among the large number of existing pegmatites developed in the earlier cycles, only a few were vulnerable to replacement or reconstitution by later cycles.

Certainly in a comagmatic province consisting of five granite cycles, four of which were extensively developed in the area, a pegmatite channelway, once established, could be reopened during successive granite invasions and could be penetrated and reconstituted by later magmatic surges.

Quirke and Kremers (1943, p. 574–580) used "pegmatite thoroughfare" to designate the space of pegmatite formation into which and along which injection of mother liquor takes place. The thoroughfare might be operative for a long time, "the process of pegmatite development, like the cycle of erosion, had its interruptions and its rejuvenations" (p. 580). The recurrent injection and replacement processes within the pegmatite thoroughfares followed channels established early in one or more granite cycles. Zones of structural weakness opened and reopened, both chemically and mechanically. Some channelways that originated in Mount Morrison times appear to be the avenues along which later pegmatite-producing materials progressed and locally came to rest during one or more subsequent granitic cycles (Table 1, nos. 9, 13).

Field observations favor the existence of at least a few pegmatite thoroughfares some of which were repeatedly activated in late Precambrian times. All the complex pegmatites are directly associated with granites of two or more cycles. The large complex pegmatites that have been enriched or replaced by later mineralizations crop out in contact with the primary "mother rock". In addition, the granitic rock of a later cycle crops out near by or is in direct connection with the complex pegmatite. Relics of primary simple pegmatite material inherited from the original simple pegmatite lie within the complex pegmatite dikes. How many times the primary pegmatite was influenced by later cycles is yet to be determined. A unit of quartz, feldspar, and fresh black biotite books with interleaved crusts of orthite enclosed in a body of pegmatite of otherwise characteristic Mount Morrison aspects is probably the result of Boulder Creek mineralization.

The factors controlling a polycyclic development of certain dikes are unknown. Probably the physical situation of a shear zone healed by primary pegmatite favored a later-cycle pegmatitization along its course. The pegmatite-filled fissures were probably rejuvenated by crustal movements, attendant on the development of bodies of granite of later cycles in a partly rigid crystalline area.

The later granite magma surges (Boos and Aberdeen, 1940, p. 723–729), near the end of the Precambrian, could have created, upon crystallization, a gigantic filter press that squeezed out the relict magmas of the earlier granite cycles into fractures of much older incipient breaks that were originally filled by early pegmatite. In this way some thoroughfares were reconstituted. This might account for the lack of similarity between the complex pegmatites.

Genesis

Answers to Questions

Field records have provided answers to many of the questions concerning the problems of pegmatite genesis in the Denver Mountain Parks area. The numbered paragraphs that follow correspond to the numbers of the questions posed under Problems of Pegmatite Genesis.

(1) Pegmatites are abundant throughout the area (Pl. 1). They outcrop within the granitic rocks, are marginal to plutons, and occur externally in the metasedimentary and metaigneous rocks enclosing the granites. Pegmatites of magmatic origin and those due to metasomatic activity occur chiefly in the vicinity of the granitic bodies to which they are genetically related. Pegmatites created by metamorphic differentiation or by ultrametamorphism occur sporadically but appear to be best developed and most numerous in the metamorphic rocks adjacent to granite. The granite-producing agents, chiefly magmatic, were the source of the energy and much of the material that developed the pegmatites.

(2) The physical, mineralogical, and structural features that distinguish the granites of one cycle from those of another are reflected in their simple primary pegmatites. For example, the marked light and dark banding and the delicate gneissosity of the Mount Morrison gneiss and migmatite is present to some extent in the pegmatites of Mount Morrison genesis. Both the gneiss and its pegmatites are quartz-poor. The trachytoid alignment of the tabular microcline twins, so conspicuous in the Silver Plume granite, occurs also in some of its late pegmatites. Some brightly-colored and spotted pegmatites are giant replicas of the spotted Pikes Peak granite. Both carry fluorite. The omnipresent biotite of the Boulder Creek granite and gneiss occurs in its pegmatites also.

(3) Most well-developed pegmatites contain sufficient reliable, intrinsic, mineralogical, structural, and other physical features to relate them genetically to a specific granite cycle. Exceptions are the well-zoned late pegmatites of each cycle that are not everywhere easily distinguished from one another unless found in the company of the parent granite.

(4) Any simple primary pegmatite that is enclosed by mica schist is more micaceous than one enclosed in massive nonmicaceous rock. Where the host is massive hornblende gneiss or quartz-mica gneiss the pegmatite is usually rich in feldspars. Pegmatites within quartzite or quartz-rich rock are quartz-rich or carry massive units of quartz. Pegmatites enclosed in granite are chiefly simple, giant-grained granite bodies. Those deep in granite show little or no enrichment or hydrothermal alteration.

In general, the aplites of all granite cycles are much alike, but the character of the host rock influences, to some extent, the mineralogy of the aplite. Aplites enclosed in hornblende gneiss are generally dense, feldspathic, flecked with schorl, and quartz-poor. Those surrounded by mica-schist and mica gneiss carry granulated quartz, fine-grained pink feldspars, and small red garnets with streaks and flakes of white mica.

(5) The earliest simple primary magmatic pegmatites of any granite cycle are massive, unzoned, and feldspar-rich. The later pegmatites of any cycle are apt to be zoned and polymineralic. The latest of all in any cycle are quartz-rich, well-defined, tabular bodies with easily recognized internal zones and sharp contacts against host rocks.

(6) The primary simple pegmatites that developed in the earliest (Mount Morrison) granite cycle differ somewhat from those developed in later cycles. In general, the Mount Morrison pegmatites are gneissic, and those of later cycles are not. Accessory minerals of both Mount Morrison and Boulder Creek cycles include considerable magnetite and orthite. The Pikes Peak pegmatites carry purple fluorite, and the Bounder Creek pegmatites have much biotite. Primary Silver Plume and Mount Olympus pegmatites are zoned, bright, fresh, rich in potash feldspar, poor in accessory minerals, and are muscovite-bearing.

No single physical, structural, mineralogical, or textural characteristic serves to establish the "consanguinity" of a primary simple pegmatite of magmatic origin and its parent granite. It is the total of all the intrinsic features of both granite and pegmatite that establish the genetic relationship when the criteria are applied in the field.

It is hoped that the criteria developed for identification of pegmatites in this area may be tested in other areas where more than one generation of granites and their related pegmatites are extensively exposed.

(7) Pegmatites of complex structure and mineralogy are usually associated on the outcrop or are in contact with two or more of the four extensively exposed granites. Many of the larger pegmatites were initiated as simple primary dikes or sills in Mount Morrison times. However, field evidence alone has not proved whether or not the conversion of a simple pegmatite into a complex one occurred entirely within Mount Morrison times or was the result of a renewal, in the area, of the granitizing and pegmatitizing processes of a later cycle. The mechanics of the reopening of a once-closed cycle are yet to be established.

(8) The writer believes that some pegmatite channels or thoroughfares were in operation throughout most of late Precambrian times, in the Denver Mountain Parks area. A few complex pegmatites that contain extensive suites of accessory or rare minerals, *viz.*, the Grover pegmatite of Table 1 (no. 14), were initiated in Mount Morrison times but were also the channels to which Silver Plume fluids had later access.

(9) The pegmatite-rich areas that carry some complex pegmatites coincide, in places, with fracture zones developed during the emplacement of the larger granite bodies of the area. Well-defined belts of northwest-trending pegmatites follow repeatedly activated shear zones and well-defined axes of major folds (Pl. 1). But some complex pegmatites are nowhere near any major structural weakness.

The field evidence and record for the complex pegmatites and their genesis cited above will be supplemented by mineralogical studies now in progress that will substantiate or modify the concepts of origins given above. The conclusions reached from field evidence have been presented here in the hope that observers of both simple and complex pegmatites in other areas, where more than one granitic rock outcrops and where there is more than one granite cycle, will look for evidence to affirm or negate the polycycle concept of the origin of thoroughfare pegmatites. It is also hoped that students of pegmatites may find evidence that will help to establish the consanguinity of granite and pegmatites where both were derived from magmatic activity engendered during a single granite cycle.

References Cited

Bastin, E. S., and Hill, J. M., 1917, Economic geology of Gilpin County and adjacent parts of Clear Creek County and Boulder County, Colorado: U. S. Geol. Survey Prof. Paper 94, p. 33–37.

Boos, Margaret Fuller, 1935, Some heavy minerals of Front Range granites: Jour. Geology, v. 43, p. 1033–1040.

—— 1936, Primary structure and mode of intrusion of some younger precambrian batholiths of the Front Range (Abstract): Geol. Soc. America Proc. 1935, p. 67.

—— 1946, Genesis of pegmatites in the Denver Mountain Parks Area, Colorado (Abstract): Geol. Soc. America Bull., v. 57, p. 1179.

—— 1949, Structure of the Precambrian, Denver Mountain Parks quadrangle, Colorado (Abstract): Geol. Soc. America Bull., v. 60, p. 1874.

—— 1951, Lime silicate and carbonate rocks of the Idaho Springs formation, Denver Mountain Parks area, Colorado (Abstract): Geol. Soc. America Bull., v. 62, p. 1533.

Boos, M. F., and Aberdeen, Esther J., 1940, Granites of the Front Range, Colorado: the Indian Creek plutons: Geol. Soc. America, v. 51, p. 695–730.

Boos, M. F., and Boos, C. Maynard, 1934, Granites of the Front Range, Colorado: the Longs Peak-St. Vrain batholith: Geol. Soc. America Bull., v. 45, p. 303–322.

Cameron, E. N., Jahns, R. H., McNair, A. E., and Page, L., 1949, Internal structures of granitic pegmatites: Econ. Geology Mon. 2, p. 13–30.

Goldschmidt, V. M., 1922, On the metasomatic processes in silicate rocks: Econ. Geology, v. 17, p. 105–123.

Hanley, J. B., Heinrich, E. W., and Page, L., 1950, Pegmatite investigations in Colorado, Wyoming, and Utah, 1942–1944: U. S. Geol. Survey Prof. Paper 227, p. 5–8, 27–30, 82–85.

Heinrich, E. Wm., 1948, Pegmatites of Eight Mile Park, Colorado: Am. Mineralogist, v. 33, p. 442–448.

—— 1953, Zoning in pegmatite districts: Am. Mineralogist, v. 38, p. 68–87.

Hietanen, Anna, 1951, Metamorphic and igneous rocks of the Merrimac area, Plumas National Forest, California: Geol. Soc. America Bull., v. 62, p. 605.

Jahns, R. H., 1946, Mica deposits of the Petaca district, Rio Arriba County, New Mexico: N. Mex. Bur. Mines and Mineral Resources Bull. 25, p. 73.

Landes, K. K., 1933, Origin and classification of pegmatites: Am. Mineralogist, v. 18, p. 33–56, 95–103.

Lovering, T. S., 1935, Geology and ore deposits of the Montezuma quadrangle, Colorado: U. S. Geol. Survey Prof. Paper 178, p. 11.

Lovering, T. S., and Goddard, E. N., 1939, Geologic map of the Front Range mineral belt, Colorado:

(accompanying text in Colorado Sci. Soc. Proc., v. 14, p. 1–48).

— 1951, Geology and ore deposits of the Front Range, Colorado: U. S. Geol. Survey Prof. Paper 223, p. 23 *et seq.*

Quirke, T. T. and Kremers, H. E., 1943, Pegmatite crystallization: Am. Mineralogist, v. 28, p. 571–580.

Spurr, J. E., Garrey, G. H., and Ball, S. H., 1908, Economic geology of the Georgetown quadrangle, Colorado: U. S. Geol. Survey Prof. Paper 63, p. 38–74.

Turner, F. J., and Verhoogen, Jean, 1951, Igneous and metamorphic petrology: p. 328–333.

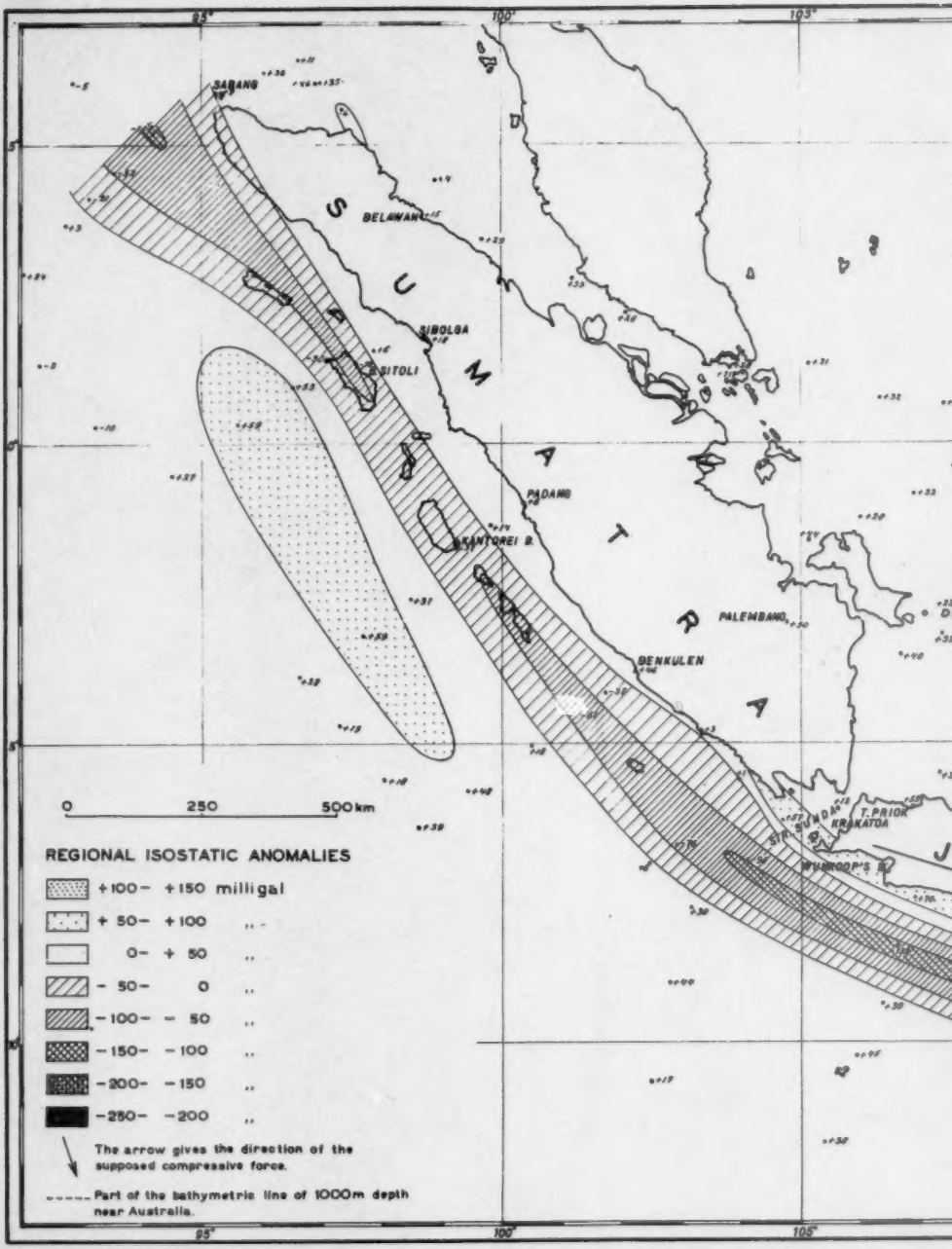
Uspensky, N. M., 1943, On the genesis of granitic pegmatites: Am. Mineralogist, v. 28, p. 437–447.

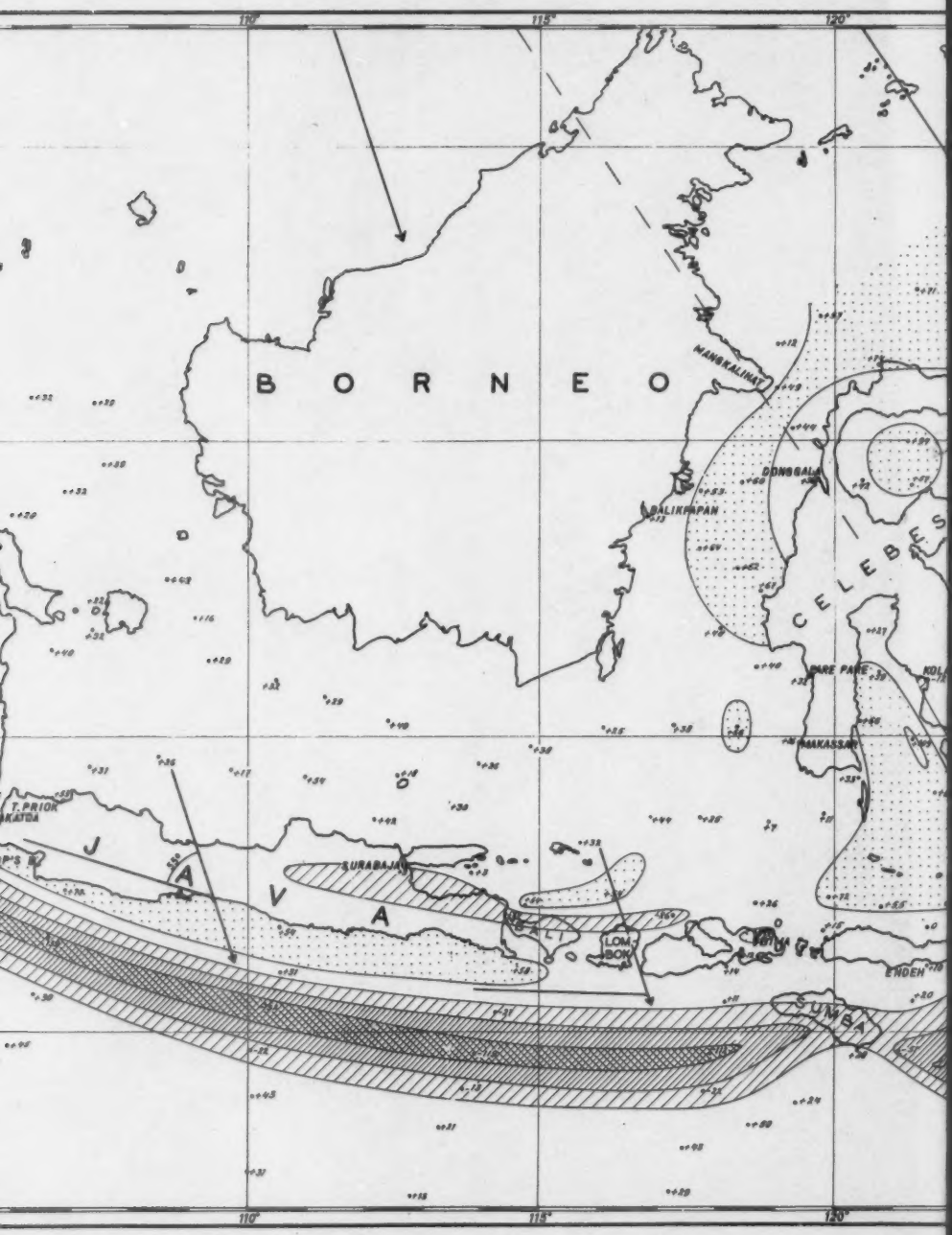
Vogt, J. H. L., 1931, Physiochemical explanation of the genesis of granites: Zeitschr. deut. Geol. Ges. v. 83.

2036 South Columbine, Denver 10, Colorado

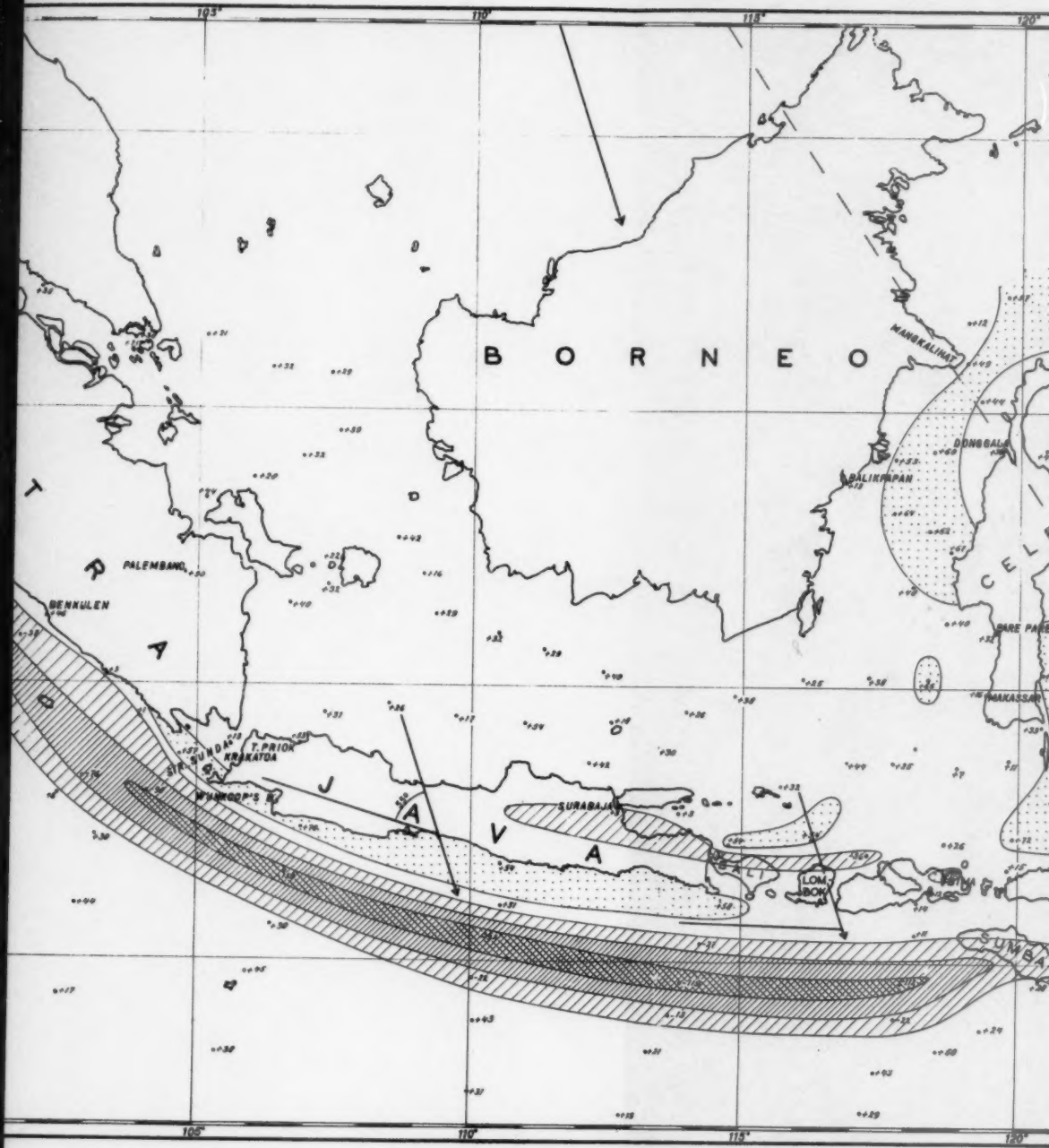
Manuscript Received by the Secretary of the Society, November 16, 1951

Project Grant 336/40.

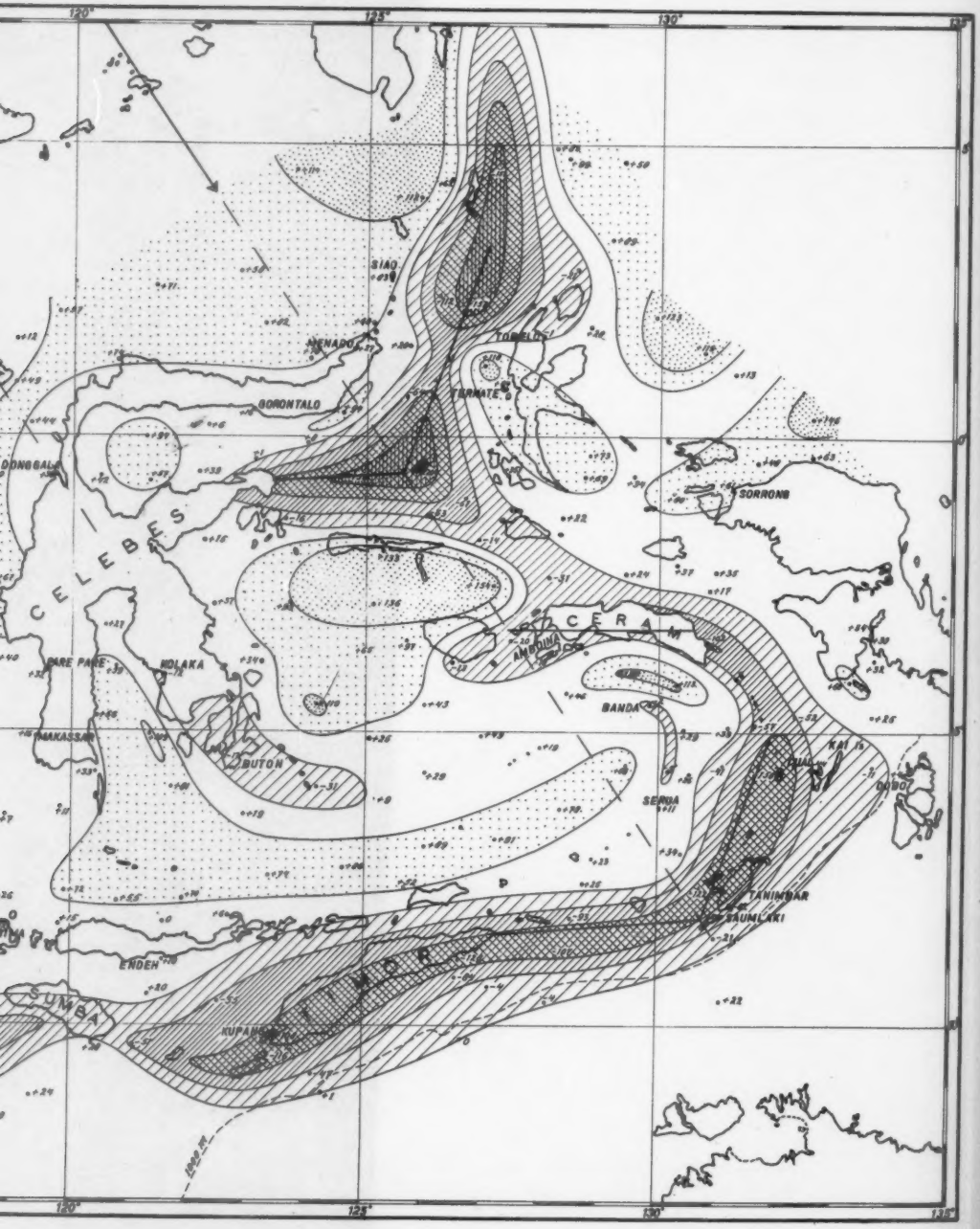

95°
100°
105°
SABANG
BELAWAN
SIBOLGA
SITOLI
PADANG
SIBERUT I.
SIPORA I.
PAGAI I.
PALEMBANG
BENGKULEN
SUNDA STR.
T. PRIOK
KRAKATOA
KRAKATOA I.
WUKOOP'S I.
SUMATRA
0 250 500 km
REGIONAL ISOSTATIC ANOMALIES
+100 – +150 milligal
+ 50 – +100 ,,
0 – + 50 ,,
– 50 – 0 ,,
–100 – – 50 ,,
–150 – –100 ,,
–200 – –150 ,,
–250 – –200 ,,
The arrow gives the direction of the supposed compressive force.
Part of the bathymetric line of 1000m depth near Australia.



GRAVITY ANOMALIES OF INDONESIAN ARCHIPELAGO



GRAVITY ANOMALIES OF INDONESIAN ARCHIPELAGO


120°
125°
130°
135°
SIAO
MENADO
GORONTALO
TOBELO
TERNATE
SORRONG
CELEBES
CERAM
AMBOINA
BANDA
MAKASSAR
BUTON
SERUA
KAI Is.
DAMAR
TANIMBAR
BAUMLAKI
ENDEH
SUMBA
TIMOR
KUPANG
1000 M.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 65, PP. 143-164, 4 FIGS., 2 PLS. FEBRUARY 1954

INDONESIAN ARCHIPELAGO: A GEOPHYSICAL STUDY

BY F. A. VENING MEINESZ

ABSTRACT

This geophysical study of the Indonesian Archipelago deals first with the belts of strong negative gravity anomalies; it studies the Archipelago also in the light of Ewing's results in the West Indies. The hypothesis of crustal downbuckling seems to cover all the facts. The main tectonic arc has been brought about by movement of a great inner crustal block south-southeast relative to the crust on the outside of the arc and, for the second tectonic arc, by movement of a northeastern block in a direction slightly diverging to the east from the first direction, relative to the first inner block.

In most places the relative block movements not only bring about a component of compression at right angles to the arc but also a component of shear in the sense of the arc. The hypothesis is advanced that this shear movement, which is strongest on the sides of the arc—*i.e.*, in Sumatra and in Mindanao—takes place in the inner arc, causing volcanic and seismic activity there and thus explaining the presence in all island-arc areas of a volcanic arc inside the tectonic one.

The writer discusses the crustal deformation in the tectonic arcs, which are supposed to have been of a plastic character, and the many shapes the surface features may assume, from a deep trough to a high mountain ridge.

In the second part, the hypothesis of convection currents in the deeper layers is discussed, first assuming currents over the whole or at least a great part of the thickness of the mantle—exerting a drag on the crust which may account for the great forces causing the relative block movements and the crustal compression mentioned—, secondly assuming currents over a depth of 500–750 km below the crust which may explain the deep and intermediate earthquake foci, the sinking of the deep basins, as, *e.g.*, the Banda basin, the Celebes Sea, etc., and the origin of a third ridge inside the volcanic arc, and thirdly assuming a still shallower type reaching down only a few hundreds of km, which may have brought about the basins of a depth of about 2000–2500 m as, *e.g.*, Strait Makassar, the Gulf of Bone, and the Gulf of Tomini.

It is usually assumed that for convection a homogeneous layer is required. Although from a depth of 900 km to 2900 km the mantle seems to fulfill this condition, it appears that from 200 km to 900 km depth the density increases too quickly to allow this assumption. We must here suppose a gradual change of phase or of chemical constitution. This point is discussed, and an attempt is made to show that a convection current can break through a layer of phase change of this kind and thus account for the origin of the basins. Also, the basins do not disappear when the currents stop.

For this hypothesis, however, one must assume that the mantle has a certain strength which must be overcome before flow or creep can set in. For this purpose a trigger effect must be supposed, brought about by a secondary phenomenon causing a horizontal temperature gradient. Griggs made this same supposition in 1939 to explain the pseudoperiodicity of the tectonic phenomena of the Earth's crust.

CONTENTS

TEXT

	Page
Introduction and acknowledgments	144
Belts of negative gravity anomalies, buckling hypothesis	144
Volcanic arc from Sumatra to the Banda Islands	154
Other tectonic and volcanic belts	157
Troughs, basins, and third ridge, convection-current hypothesis	158
Idiogeosynclines	163
References cited	163

ILLUSTRATIONS

Figure	Page
1.—Profiles through the Alps	146
2.—Shear pattern of the Earth's crust, caused by shift of the poles	153
3.—Maps of foci of 0–50 km depth, 50–100 km depth, and 100–150 km depth	156
4.—Vertical cross section of convection current	160

Plate	Facing page
1.—Gravity Anomalies of Indonesian Archipelago	143
2.—Contour Map showing Basins of Indonesian Archipelago	164

Introduction and Acknowledgments

This paper summarizes geophysical studies about the Indonesian Archipelago, partly referring to older papers and results, partly to more recent and new ones, and taking into account new data, among others those provided by Ewing's recent research. It is a privilege that it can appear with Ewing's paper about his important results in a similar island-arc area.

The writer is conscious of the hypothetical character of much of what follows. The problems dealt with refer to the phenomena occurring or having occurred in the Earth's crust and in the deeper layers of the Earth. The data we dispose of are mostly given by geology, seismology, and gravimetry. For the geological features the writer uses mainly the expositions given by Umbgrove about the Indonesian Archipelago. He also often refers to the work of Hess, Kuenen, and others.

Belts of Negative Gravity Anomalies, Buckling Hypothesis

The two belts of negative gravity anomalies in the Indonesian Archipelago (Pl. 1) coincide with the outer arcs. Long parts of these arcs, however, do not come up above sea level. The first belt runs over the islands west of Sumatra and is probably the continuation via the Andaman and Nicobar islands of the belt observed by Evans and Crompton in Burma; it goes on over a submarine ridge south of Java, over Timor, the Tanimbar islands, the Kai islands, and Ceram. It is accompanied by an inner arc of mostly volcanic character over Sumatra, Java, and the smaller Sunda islands up to the Banda islands. Northwest of Ceram this first negative belt has a weak connection with the second belt, less regular in its intensity, which runs over East Celebes, the Molucca Sea, and the Talaud islands toward the deep Philippine trench. The corresponding volcanic arc is found over North Celebes, the Sangihe islands, and Mindanao.

These belts of negative anomalies coincide with all the islands and areas where strong folding and overthrusting of the surface layers has been found. This points to crustal shortening by lateral compression of the crust in these belts during the last period of strong tectonic activity. According to Umbgrove this can probably be dated in the Late Miocene—*i.e.*, in Tertiary f 2 when using the division of the East Indian Tertiary given by Van der Vlerk (Van der Vlerk and Umbgrove, 1927; Van der Vlerk and Leupold, 1931).

The coincidence of the areas of strong folding and overthrusting with the belts of negative anomalies suggests that the deficiency of gravity has been caused by lateral compression of the crust. This has brought the writer to the hypothesis that the deformation of the crust in those circumstances does not lead to a thickening during which isostasy is maintained, but to a downbuckling of the crust which brings it to a much lower position than would correspond to isostatic balance. It is clear that this must give rise to strong negative anomalies in the downbuckled belt. We shall presently go further into the question of the mechanism of this phenomenon and we shall see that such a behavior of the crust can readily be explained. We thus can also understand that the first stage of orogeny is the development of a geosyncline. In all cases where sediments are provided by neighboring areas sedimentation must have favored the sinking of the belt.

As long as the compression works we may suppose that, because of the buckling phenomenon, the buoyancy of the belt cannot lead to the local readjustment of the isostatic equilibrium, but it must lift up the whole crustal area, and so we may expect positive anomalies over the adjoining belts. The anomaly map of Indonesia shows evidence nearly everywhere that this is the case.

When, however, in a later stage of the tectonic phenomena the compression lessens, a beginning of isostatic readjustment may be expected, leading to a local rising of the downbuckled belt and a falling back of the adjacent areas. We may probably assume that this stage is reached in the eastern half of the Indonesian archipelago where the islands in the main tectonic belt—as, *e.g.*, Timor—show a quick rising while to both sides troughs are developing, separated from the islands by faulting or flexures.

In mature folded mountain ranges where the isostatic balance is nearly restored—as, *e.g.*

the Alps—this post-folding rising has led to high elevations. A remnant of negative anomalies in the central area of the past orogeny is, however, still present. We need not enlarge here on the last stage of the phenomenon, the disappearing of the mountain system by erosion. We see that the buckling hypothesis fits the broad lines of the history of geosynclines as it has long been known to the geologists.

In detail, however, there are great differences among geosynclines, and this is likewise true for the ways in which the mass deficiencies in the downbuckled belts occur. If a crust consisting of a sedimentary layer, a granite layer, and an "intermediate" layer is pushed together we may expect that during buckling the two lower layers will thicken and will bulge downward, the granite layer probably forming a smaller downward bulge than the lower layer. In many cases, however, the sediments, being less competent than the deeper layers, have not been carried along downward but have been pushed in more or less horizontal folds over the surface, as it has, *e.g.*, been especially for the Alps. Or we may say that the lower crustal layers are pushed below the central compressed mass of the sedimentary layers.

In Argand's and Staub's well-known geological profiles through the Alps we see clear evidence of the zone of downbulging of the crust. In the three profiles given in Figure 1 we find them in the areas of the Hohe Tauern, Campo Viola, and Monte Rosa Nappe. On both sides we see fault planes in the deeper layer tilting downward toward the central zone with wedge-shaped structures between.

For other mountain ranges, as, *e.g.*, the Pyrennees, the folds and overthrusts have had a less extensive development than in the Alps, but we find many similar wedge-shaped structures with fault planes tilting in the same way toward the central belt and pointing to the same phenomenon of crustal downbuckling, in this case though with less sediments pressed out. In a paper on *Pliocene uplift of Tertiary mountain chains* L.U. de Sitter (1952) considers this an indication that no large roots could have developed there and that, therefore, the uplift in these areas could not be attributed to isostatic readjustment; this objection would at the same time invalidate this explanation for the Alps. As, however, the folds and overthrusts at the surface are no measure for the shortening and downbulging of the crust, this objection does not seem valid.

We shall not further pursue here his discussion of the geological surface features and shall now take up the question of the distribution of the mass deficiencies caused by the downbuckling. We may expect them in general where lighter layers have been pushed downward and in these "roots" replace deeper layers of greater density. They must be proportional to the density difference and to the cross section of the roots, and we must expect these roots to be larger for deeper discontinuity surfaces. The sum of the density differences must equal the difference of the densities of the plastic substratum and of the surface layer of the crust.

In the practice of the isostatic reduction of gravity values according to the Airy-Heiskanen method for local compensation and according to that of the writer for regional compensation, this picture, for the sake of simplicity, is usually replaced by the schematic one which assumes the crust to consist of only one homogeneous layer of a density of 2.67 floating on a substratum of a density of 3.27. So, in our case, we would have only one root of a density deficiency of 0.6 at the lower boundary of the crust. But we must never forget—and the writer has often pointed it out (Vening Meinesz *et al.*, 1934; Vening Meinesz, 1941)—that this replaces the more complicated situation as described. In most cases we shall at least have a root between the sediments and the granite, between the granite and the intermediate layer, and between the intermediate layer and the plastic substratum. It is possible, moreover, that the boundary between plasticity and rigidity is not sharp. This boundary may perhaps also be found within one layer, and this would imply only a small density difference or none at all at the lower boundary of the rigid crust.

These considerations are valid not only for the roots which, according to the Airy view of isostasy, we may expect below topographic elevations, but also for the roots brought about by the supposed downbuckling of the crust which we are here dealing with and of which the mass deficiencies are not at all or not suffi-

FIGURE 1.—PROFILES THROUGH THE ALPS
A after E. Argand, B and C after R. Staub.

ciently compensated by surface topography. We thus come to the viewpoint that the way in which the mass deficiencies caused by the crustal downbuckling are distributed may show great differences, although for a certain amount of downbuckling the total mass deficiency must be the same. This last statement becomes clear if we realize that for studying the buckling mechanism of the crust in its great lines we can consider the crust as a floating plate on which at each point of the surface a force works corresponding to the buoyancy caused by the total mass deficiency in the vertical column of the crust at that point.

We thus apply the physical interpretation of the gravity anomalies (Vening Meinesz *et al.*, 1934, p. 23), which has seldom been applied. Without especially investigating the distribution of the masses bringing about the anomalies —we shall call this the geometrical interpretation—we study the deviations from equilibrium and the forces bringing them about. The two interpretations cannot be entirely separated. If in a vertical column of the crust a certain mass deviation is present causing a disturbance of balance, the anomaly caused at the surface depends on the depth of the mass, but, for larger horizontal extent of the mass, this uncertainty becomes less.

As the horizontal cross section of the belts of negative anomalies is a few hundred km and the maximum intensity more than 100 mgal it is probable that this means a disturbance of the balance of the crust which would disappear relatively quickly, probably in less than 100,000 years, if it were not maintained by forces working in the Earth. For this purpose it is likely that we have the choice between lateral compression of the crust or a disturbing phenomenon in the substratum. We shall see that the first is the more probable explanation. So this would lead us to the same conclusion as we had already arrived at because of the folding and overthrusting found in the islands in the belt.

We shall now return to the different possible constitutions of the crust discussed above and examine the Indonesian archipelago. Nearly everywhere the belt coincides with a ridge; where it rises above sea level folding and overthrusting have been found. This points to the presence of a sedimentary layer which has not, or not entirely, been carried along downward during the downbuckling. It is likely, therefore, that we have here a more or less normal crust of continental character with a downbulged root at each discontinuity surface and also one at the lower boundary of the crust.

South of Java the situation is different. It appears that there the sediments do not fill the trough entirely; we find a trench south of the submarine ridge. The ridge coincides with the belt of negative anomalies. As the sediments have probably for the most part come from the north side here, we can understand that, if they were insufficient to fill the trough, a trench was left on the south side.

East of the Philippines the deep trench coincides with the belt of negative anomalies. In the free-air anomalies the belt shows strong negative values, but the regional isostatic anomalies with a radius of distribution of the compensation R of 116.2 km show the belt only slightly, and in the curve of anomalies for R = 232.4 km we do not see it at all. We may conclude that the belt of negative anomalies can be entirely explained here by the deficiency of matter as given by the topography. This result, however, is founded on only one gravity profile over Strait Surigae, north of the island of Mindanao.

We know much more about the Puerto Rico trench where also a belt of negative anomalies coincides with the trench. Here the spectacular seismic results found by Ewing and reported in his paper show the presence of a layer of unconsolidated sediments of a thickness of many kilometers whose bottom could not be detected seismically. From the gravity values he derives that a thickness of this layer of 6 km, combined with a 3-km layer of density 2.67 above the ultrabasic layer of a density of 3.27, could explain the anomalies. This leads to the conclusion that probably the rigid crust exists here for a great part of ultrabasic material which is overlain by only a thin layer of a few km of sialic or simatic matter. This would make it similar to the crust under the deeper ocean basins where recent seismic results show that under a sedimentary layer which may be quite thin a layer is found of some 3–6 km showing the

seismic velocity of basalt. Below that is the main crustal layer where the seismic velocity points to ultrabasic matter.

We see that Ewing's results show us the remarkable fact that the rigid crust in the trench (coinciding with the belt of strong negative anomalies) has sunk to a depth of the order of 15 km. In the light of what has already been said the writer thinks that we may interpret this as a spectacular instance where the downbuckling of the crust has directly been detected. In this case where the main part of this crust has nearly or entirely the same density as the plastic substratum, the entire density deficiency corresponding to the gravity anomalies is found at the surface of this rigid crust—*i.e.*, in the unconsolidated sediments and in the topographic deficiency of matter.

Long ago Hess (1939) expressed a similar idea in explaining that in the southern part of the Antillean belt of negative anomalies the downbuckling of the crust is covered by sediments pressed out and leading to a ridge showing the islands of Barbados, Tobago, and Trinidad, but that this is not the case in the northern half where the continents are too far away to provide sediments to form a ridge.

We have not yet examined another result of great importance of Ewing's seismic research—*viz.*, the fact that he found the rigid crust north and south of the Puerto Rico area to have oceanic character. Below a thin layer of sediments and a layer of basaltic velocity only 3 or 4 km thick he observed wave velocities of 8 km/sec which correspond to ultrabasic rocks. Only in the Puerto Rico area an upper layer of a thickness of about 10 km was detected showing the lower velocities of sialic rocks. It is hardly necessary to point out the paramount significance of this result for the problem of the origin of granite; it is very unlikely that originally a narrow belt of sialic matter was present here.

We shall not attack this great and difficult problem here; it does not form a central part of our subject. The writer, nevertheless, cannot resist the temptation to put two questions here without, however, attempting to give an answer to them. (1) Is it perhaps likely that many geosynclines have started in this way by the development of a deep trench in an area where the crust consists mainly of ultrabasic rocks? (2) Have the sediments accumulated in these trenches, at least partly, been carried downward during the further downbuckling of the crust, and under the effects of high pressure and of high temperature could they have been transformed into granite and as such been erupted at the surface of the crust? It is well known that such eruptions of sialic magma have frequently occurred during later stages of the orogenetic phenomena.

We shall now consider the pattern of the belts of negative anomalies in the Indonesian archipelago and examine their intensity. In the first place we are struck by the linear shape and the consistent pattern of the first belt over the whole archipelago from west of Sumatra via Timor and the outer Banda arc to Ceram and Buru; it shows only a short irregularity south of the island of Sumba. This makes it unlikely that the belt is caused by chemical or differentiation phenomena in the substratum; we can hardly imagine how such phenomena in these plastic layers could assume a linear and smoothly curved shape and bring about such a regularly varying intensity of the gravity anomalies. With crustal downbuckling these features can on the contrary be readily explained; we shall presently find this confirmed when studying its mechanism. This will give us a strong argument in favor of our buckling hypothesis.

An important question arises regarding the type of compression of the crust which brought about this deformation: Is it all-sided or has it one direction? The way in which the intensity of the anomalies in the belt varies throughout the whole archipelago clearly points to the second supposition. We can well understand this distribution by assuming that this first belt of crustal downbuckling has been caused by the relative movement of two great crustal blocks in such a way that the inner one, covering nearly the whole archipelago, has moved to the south-southeast relative to the one on the outside of the belt—*i.e.*, in the direction of the arrow on the anomaly map (Pl. 1). Such a movement would explain the great intensity of the anomalies of the belt in the part where it makes a great angle with the direction of the compression—*e.g.*, south of Java and over the islands

from Timor to the Tanimbar group. It likewise makes clear why the anomalies are weaker in those areas where the component of the movement at right angles to the belt is small, as is the case in the belt west of Sumatra and in the part between the Kai islands and Ceram. In this case the crustal shortening in the belt by the relative movement of the two blocks must have been much smaller.

For the origin of the second belt of negative anomalies from East Celebes via the Molucca sea and the Talaud islands to the Philippine trench we have to assume a third block north and west of this belt which is probably separated from the inner block mentioned above by a zone of a certain breadth running over Northeast Borneo and the southern part of the north arm of Celebes (*see* the hatched line in Plate 1); it is further limited by the above second belt. This block may be assumed to move with regard to the inner block in the direction of the arrow in that area on the anomaly map—*i.e.*, slightly diverging from the direction of movement of the inner block. We see that over the Philippine trench this direction makes only a small angle with the belt, and this is again in harmony with the smaller anomalies here.

We find a strong confirmation of our hypothesis in the fact that Sumatra and Mindanao show major faults parallel to the axis of these islands and to the adjacent belts and that in Sumatra the crustal part Northeast of these faults during the earthquake of 1892 has been proved to move to the Southeast with regard to the block on the Southwest side (Muller, 1895). This relative movement parallel to the belt constitutes the second and greatest component here of the total relative movement, parallel to the arrow, of the inner and outer blocks. For Mindanao the relative movement during earthquakes on both sides of the faults seems also to have been in harmony with our hypothesis; the southwest side has moved relatively to the northeast. So the hypothesis of relative movement of the three crustal blocks is probable not only because of the folding and overthrusting in the tectonic belts between these blocks but also because of the crustal movements during earthquakes found in Sumatra and in Mindanao. We thus also obtain an argument in favor of the supposition that folding and overthrusting in tectonic belts are indeed brought about by a shortening of the whole crust.

The fact that the second component of the relative movement which is parallel to the tectonic belt takes place along faults in Sumatra and in Mindanao—*i.e.*, at some distance inside this belt and not in the belt itself—is an important point which we shall discuss later. We may now draw attention to the fact that the faults thus coming into action are situated in the volcanic zone of these islands, which forms part of the volcanic arc of the archipelago. This suggests that the volcanoes may have formed here because the relative movements of the great crustal blocks caused faults through the whole crust in this zone. We shall come back to this hypothesis.

We shall now attempt a theoretical discussion of the type of deformation we may expect the crust to undergo with lateral compression. For two reasons it appears probable that this deformation is plastic (Vening Meinesz, 1950). In the first place the buckling limit of the compressive stress for a rigid crust of a thickness of 30 km is of the order of 40,000 to 50,000 kg/cm^2 or higher; it seems far too much to assume that the crust reacts elastically up to that limit. Below that limit and as long as they are in the elastic stage, the crustal layers must remain horizontal, and so no perceptible anomaly can arise. It is true that by assuming the crust to consist of different layers which can glide over each other without too much friction, we should find a lower buckling limit, but for bringing this limit down to an acceptable value—*e.g.*, 3000 kg/cm^2—we should have to admit 15 layers all moving without mutual friction over each other, and this seems likewise inadmissible.

A second reason for assuming plastic deformation in the tectonic belt from the start of the phenomenon was indicated in 1936 by Bylaard, who pointed out that the beginning of buckling as an elastic deformation of the crust ought to occur in a belt at right angles to the direction of the main compressive stress. If we assume this direction to be parallel to the direction of the relative movement of the crustal blocks on both sides of the belt, we see that in the greater part of the belts in Indonesia this is not the case. For the beginning

of a plastic deformation, on the contrary, this is different. Bylaard has derived the following formula for the angle φ between the direction of the main principal stress and the belt of deformation. For this case

$$(1) \qquad \cos 2\varphi = -\frac{\rho_1 + \rho_2}{3(\rho_1 - \rho_2)}$$

in which ρ_1 and ρ_2 are the principal stresses working in the plane of the crust—*i.e.*, on cross sections of the crust at right angles to it and to each other; ρ_1 represents the main compression working in the crust, ρ_2 is supposed to be zero in our case. For this last assumption we find φ to be about 55°.

If we accept the compression to have the same direction as the relative movement of the blocks it is interesting to see on the map in how many places this angle occurs in the Indonesian Archipelago. With the exception of the borders of the belts west of Sumatra, east of Mindanao, and between the Kai islands and Ceram, nearly the whole course of the belts can thus be explained. This seems to be a strong argument in favor of the assumption that the crustal deformation must have been plastic from the start.

The writer (1950) investigated the problem of what type of plastic deformation the crust must be expected to undergo as a result of lateral compression. It is clear that the compression has to exceed the limit of elasticity, often called the strength, of the crustal layers; below this limit the crust remains flat, and the elastic deformation is limited to a small elastic shortening combined with a slight thickening. If, however, the elastic limit is passed in the weakest part of the crust where this limit is first reached, plastic deformation sets in, and a nonelastic thickening of a permanent kind takes place.

At the same time another thing occurs of great importance for the further development of the phenomenon; the thickened part of the crust continues to adjust its isostatic equilibrium, and, as the viscosity coefficient of the crust must be much larger than that of the substratum, we may expect this readjustment practically to keep pace with the plastic thickening of the crust. For this to be true we need not assume the crust and the substratum to react as Newtonian liquids or, in other words, to have a constant ratio of deformation speed to stress.

As the crustal density is much more than half that of the substratum, the result of this adjustment must be that the upward bulge of the crust is smaller than the downward bulge, and consequently the compressive stresses in a crustal cross section can no longer be evenly distributed if the resultant has to keep in its normal place; the stress at the upper limit of the crust must become larger than at the lower boundary. This must bring about a down bending of the crust which must increase the stress asymmetry in the cross section and so the bending must accelerate and lead to an entire downbuckling of the crust. For details of this process the reader may refer to the writer's paper on this subject.

In connection with Ewing's results the question has to be dealt with whether such a buckling of the crust could also be explained if the crust has the constitution as found by Ewing for the Caribbean. It appears practically sure that this is indeed the case. The above reasoning applies not only to the thin surface layer of basalt or gabbro which by the lateral compression must lead to a small stress asymmetry in the crust, but it also must apply to the main crustal layer of ultrabasic matter. Even if this has the same chemical composition as the subcrustal layer, we must suppose that, no doubt because of its lower temperature, it offers more plastic resistance to deformation than does the deeper layer. By compression this surface layer must increase in thickness and behave in the same way as a crustal layer of lesser density would. The temperature conduction may be assumed to be too slow for sufficiently heating the lower part of this downbulging layer to prevent the stress asymmetry described above to come also into being in this case. So we may in the same way expect a downbuckling of the crust as a result of the lateral compression.

We have already discussed that in this case the entire mass deficiency causing the negative anomalies in the belt must be found at the surface of the crust. If the trench formed by the downbuckling ultrabasic crust contained no sediments the negative anomalies of the belt

would disappear by applying the topographic reduction. If the trench is of the type of the Puerto Rico trench and would be partly filled by the unconsolidated sediments as discovered by Ewing, the anomalies would diminish by the application of the topographic reduction, but they would not disappear.

It is clear that if the subcrustal ultrabasic layer has a slightly greater density that that of the crust a corresponding part of the negative anomalies would be caused by the mass deficiency represented by the crustal root.

So far we have assumed the crustal deformation to have been mainly caused by lateral compression of the crust. For the belt west of Sumatra, east of the Philippine Islands, and for the area between the Kay Islands and Ceram this is certainly not the case. Here the principal phenomenon must have been a relative shearing movement of the crustal blocks on both sides of the belt in a direction parallel to it, while the compression at right angles to the belt must have been comparatively slight. In the Caribbean area probably also parts of the belt have this character, but we shall not deal with them here; we shall limit our discussions to Indonesia.

It seems likely that in the beginning of the deformation the shear movement occurred in the main tectonic belt and that only in a later stage, when the crust in this main belt had become stronger because of the thickening by compression, did the shearing take place in a parallel zone at the inside of the main belt and probably is still taking place there. For the Sumatra area this shear belt no doubt coincides with the volcanic zone in the west of the island.

For dealing with this shear movement we have to apply entirely different ideas from those discussed for the compression of the crust. It seems indicated that we should here follow the theory of shear as it was applied to geology by Chamberlain and Shepard and recently further elaborated by E. M. Anderson (1942) and M. King Hubbert (1951). They started from the development of shear planes in loose sand where experiments and theory are in good harmony about the originating of shear planes parallel to the middle principal stress and making an angle of $45° - \frac{1}{2}\varphi$ with the direction of the largest principal stress. In this formula φ may be called the angle of internal friction of the material. It is given by the formula

$$tg\varphi = \frac{\tau}{\sigma} \tag{2}$$

where τ and σ are the shear stress and normal stress components working in the shearing plane at failure. The quantity φ is the maximum angle of repose of loose sand.

It can be shown that for cohesive materials the same theory holds true. At failure shear planes come into being, parallel to the middle principal stress and making an angle of $45° - \frac{1}{2}\varphi$ with the direction of the largest principal stress. The angle φ is again the angle of internal friction; in this case it is given by

$$tg\varphi = \frac{\tau - \tau_0}{\sigma} \tag{3}$$

in which τ_0 is a cohesive shear stress.

The size of the angle φ is a function of the material; Hubbert (1951, p. 365) mentions, as the result of many experiments made for determining it, a value of $28° \pm 4°$. In rocks this would, therefore, lead to reverse faults—*i.e.*, faults made by compression—making an angle of $45° - \frac{1}{2}\varphi = 31° \pm 2°$ with the direction of the compression.

If we may again assume that the direction of the compression approximately coincides with the sense of the relative movements of the crustal blocks as supposed in this paper, we see that the direction of the anomaly belt in the border areas west of Sumatra and east of the Philippine Islands, where we suppose the shear to take place, is in good harmony with this formula. In both parts the angle is somewhat smaller—*i.e.*, 25°, but it may be mentioned that in Hubbert's sand experiments this same slightly smaller angle was found; he explained it as an effect of friction. Moreover, the directions assumed in this paper for the relative movements of the great crustal blocks can hardly be more accurate than within a few degrees.

For applying this theory here in this way we must show that the middle principal stress occurs in the sense of the vertical, at least in the beginning of the crustal deformation when the fault has come into being. This does not seem

unlikely if we assume that the pressure caused in the crust by gravity is not entirely distributed hydrostatically but that the vertical stress is slightly predominant over the horizontal.

The component of the relative movement at right angles to the belt in the Sumatra and Philippine areas no doubt brings about an asymmetric downbulging of the crust with the ocean side lower than the island side; we may perhaps even call it an overriding of the latter over the former. Still, the gravity anomalies show evidence of a slight downbulge too. Southwest of the Sumatra belt and farther seaward than the trough bordering on the belt the ocean floor shows a slight and broad rise marked by positive gravity anomalies and probably representing an upward wave of the crust accompanying the trough caused by the downbulge and by the overriding by the crustal block on the northeast side of the belt.

In these border parts of the belts, west of Sumatra and east of the Philippine islands, the relatively weak negative anomalies in the tectonic belt show that the crustal shortening in a sense at right angles to it is much less than in the middle part of the Indonesian arc. So our hypothesis made in the beginning—that the varying intensities of the negative anomalies in the different parts of the two belts in Indonesia point to two great blocks, with a third in the northeast, moving relatively in the sense of the arrows of the map—can no doubt be maintained.

We have seen that there are fairly strong indications that the direction of the main compression in the crust coincides more or less with the sense of the arrows. These indications are found in the direction of the border parts of the arc, west of Sumatra and east of the Philippine Islands, where the belt makes angles of about 25° with the sense of the relative block movements, and in all the parts of the main arcs where lines on the map indicate that the belt makes angles of 55° with these relative movements.

Even the slight curving of the belt south of Java up to the island of Sumba can probably be explained. In the beginning the deformation in the belt implies a component of movement to the east in the sense of the belt's axis, and this may be expected to have caused a slight secondary stress in this same sense increasing from a value zero in the western part to a certain value near Sumba. Applying formula (1) we see that this must bring about a slight increase of the angle of 55° from west to east thus causing the small curvature. For details of this explanation the writer may refer to his paper of 1950 (p. 44).

Before leaving this subject the writer points out a curious coincidence which may have a deeper meaning. In 1943 he published a possible hypothesis for explaining the occurrence over the Earth's surface of preferred directions in the topography as they had already been discovered and described long ago by several "lineament-tektoniker" such as Daubrée, Sederholm, Sonder, Hobbs, and others. He suggested explaining them by the supposition that the Earth's crust was displaced with regard to the poles of the Earth's rotation—*e.g.*, by a movement of the crust over the Earth's interior. This displacement, if it occurred, probably must have taken place during an early period of the crust's history. Because of the Earth's flattening it must have caused strong stresses in the crust leading to shear, and this must have left a pattern of zones of weakness in the crust for which the writer has derived formulas; it could be brought to a satisfactory coincidence with the existing preferred directions of the topography. The result is represented in Figure 2. There are indications that these zones are fairly broad. For details the writer refers to his papers on this subject (1943; 1947a). There is remarkable agreement between this pattern and the directions of relative block movements in Indonesia and in the Caribbean area; even the slight divergence of the two arrows in our Indonesian map can perhaps thus be explained.

If this coincidence is not fortuitous, it might be interpreted as an indication that the relative block movements discussed in this paper are caused by old zones of weakness of the crust giving way to forces working on the crustal blocks, the zones being broad enough to allow deviations of direction of 25° over the whole length of Sumatra without overstepping the border. If our interpretation is true it seems that the resulting compression in the Indonesian blocks has the same direction as the relative

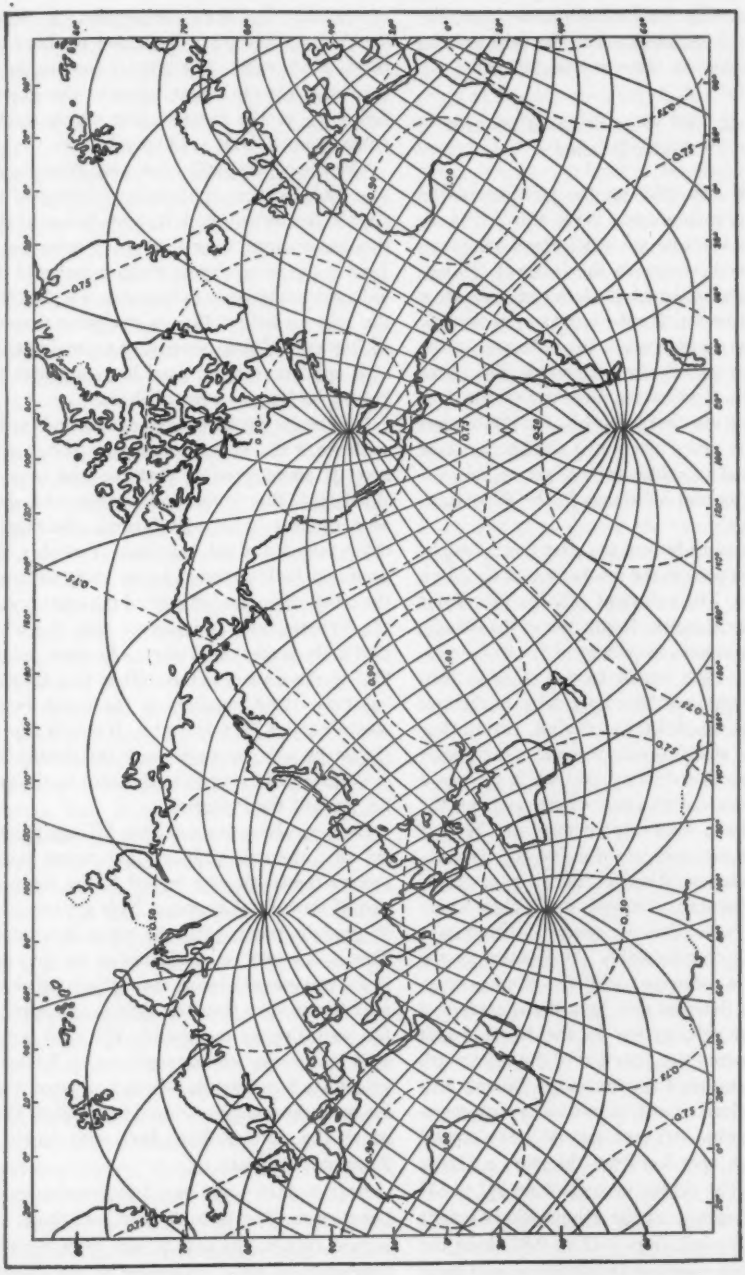


FIGURE 2.—SHEAR PATTERN OF EARTH'S CRUST CAUSED BY SHIFT OF THE POLES
After Meinesz (1947a)

movements. We shall refrain here from discussing any hypothesis about the origin of the forces working on these crustal blocks.

Volcanic Arc from Sumatra to the Banda Islands

We shall now take up the problem of the volcanic arc in Indonesia. As is well known we find such a volcanic arc in all island-arc areas. Everywhere it occurs on the inside of the tectonic belt, and, although the latter may not show up by positive topography or may even coincide with a deep ocean trough, the volcanoes of the volcanic arc practically everywhere rise above sea level. So, if we have a single arc system, it is the volcanic arc that is visible. On the borders of the arc in many places it is straight for some distance and coincides with larger islands, as, *e.g.*, Sumatra and Mindanao in the Indonesian Archipelago.

We have already seen the part these straight border belts play in the mechanism of the block movements. The principal relative movement here is shear along the length axis of the islands, and only a small component of it causes compression at right angles to the tectonic belt. We have supposed this compression in the belt strengthens its resistance against shear, thus causing the shear to take place in a belt inside the tectonic one. It may be that it chooses a belt of older deformation where perhaps the crust has been weakened; it may also be that other circumstances are decisive for its location. For Sumatra it seems difficult to say more than that the shear obviously takes place where the faults occur that are active during earthquakes and that probably the location of the volcanoes is connected with the shear zone.

Between Sumatra and Java the character of the crustal deformation in the tectonic belt changes as does the direction of the belt; south of Java it makes a greater angle (about 55°) with the relative block movement which corresponds to what we may expect if the crustal deformation here has from the start a plastic character. The change of angle from 25° to 55° means an increase of the compression at right angles to the belt from 0.42 to 0.82 times the relative block movement and a decrease of shear parallel to the belt from 0.91 to 0.57 times this movement. So the compression is about doubled, and the shear diminishes to about two thirds of its value. The greater compression is confirmed by the larger value of the gravity deficiency in the tectonic belt which reaches values here of about −130 mgals.

The shear must still have been considerable here and, in view of the strengthening of the crust in the tectonic belt, it seems indicated that this has occurred over the row of volcanoes of Java in the same way as we have assumed this to have been the case in Sumatra. The fact that the volcanic belt of Java is straighter than the slightly curved strip of negative anomalies south of it appears to be in good harmony with the idea of its being a zone of shear.

In Western Java we may perhaps find further evidence of the mechanical picture as it is given here. It seems possible that the east coast of Wynkoop's Bay is the prolongation of a shear zone running from near Jacarta (Batavia on older maps) toward this bay. Probably this zone may be interpreted as the cross section of the crust where the full shift of the crust toward the tectonic belt, as it occurs over the whole part of Java east of it, starts. We must assume that to the west of it the shift in this direction has been much smaller, as the break in the south coast of Java indicates. If this is true we should be able to understand the double belt of volcanoes running from this area to the east, *i.e.*, toward the Priangan.

Long ago Brouwer and Smit Sibinga pointed out that the axis of West Java seems shifted toward the south with regard to the continuation of the axis of Sumatra. This appears to be in good harmony with the views here given. This shift might be considered as the first part of the movement toward the tectonic belt while near Wynkoop's Bay the shift is increased by the part already mentioned. We need not be surprised to find the strongly active Krakatoa volcano in Strait Sunda, which is situated where the main shear zone is crossed by the fault along which the shift of West Java with regard to Sumatra took place.

East of Java some new features change the phenomena. The tectonic belt here shows two inward curves: the first is near Sumba and is characterized by a weakening of the negative anomalies, and the second is over Timor and

some small islands to the east. The two inward bendings are separated by an outward bending. As the map shows the directions of the belt from Timor to the Kai islands and further to Ceram can well be explained by the angle of 55° with the sense of the compression which corresponds to plastic deformation in the tectonic zone. An older explanation accounts for the shape of the belt here by assuming that it adapted itself to the border of the Australian continent. The question may be raised whether the first explanation would imply that the Australian continent has no border here or whether the area of the Banda basin and possibly also of the basins farther north have continental character. The geological evidence as given *e.g.* by Umbgrove seems to favor the latter hypothesis.

The direction of the axis of Sumba appears rather to indicate crustal shear in the same way as the belt west of Sumatra. The weakening of the negative anomalies and the volcanic activity in this island during the Tertiary seem to be in harmony with this hypothesis.

We can easily understand that in an area of such variable directions of the tectonic belt the movement of the inner block toward this belt must everywhere give rise to shear stresses, and thus a volcanic arc can be explained here in the same way as in the Java area. It seems interesting to see that this latter arc seems to continue here in a straight line and extend farther east along the general curve around the Banda Sea without deviationg as does the tectonic belt. On closer examination, however, we notice (1) that the islands north of Timor formerly showed volcanic activity but that now this is no longer the case, and (2) that we find four volcanoes in the Banda Sea—*viz.* Gunung Api, two submarine volcanoes of the Tukang Besi Is., and Pulu Komba—which could perhaps be supposed to play the part of the volcanic arc in this period. In the same way, north and northwest of Sumba the line of volcanoes of Tambora and Pulu Sangean bends northward slightly. In both areas the number of volcanoes is abnormally small while north of the outward bend, on Flores and the Solor Islands, we find an exceptionally large number.

We may perhaps attribute this to an effect of the curvature of the tectonic belt. This could be brought about by the fact already assumed that we may probably expect the movement of the crustal block near the tectonic zone, because of the downbulging of that belt, to take place at right angles to the belt, as well as in curved parts of it. This must bring about a tensile stress or, at least, a lessening of pressure in the crust in the sense of the belt on the concave side of the belt. On the outside of the curve the crust must on the contrary be under extra pressure in this sense. We can thus explain the general fact, of which an instance is mentioned above, that volcanoes are found especially on the inside of a curved tectonic arc but seldom on the outside of the curve. The above-mentioned varying density of volcanoes in the arc between Java and the Banda sea may well be explained in this way.

According to the views here developed the presence of a volcanic arc may be attributed to the fact that in general the tectonic arc is not perpendicular to the sense of the compression but that because of the plastic character of the deformation it must at the start of the phenomenon make another angle with it, often an angle of 55°; if the main character of the deformation is shear, the angle is 25°–30°. We may further expect that during the later development of the deformation, when the crust has started downbulging, shear in the sense of the zone is no longer possible, and, therefore, the component of the crustal movement parallel to the tectonic belt must occur at some distance from it and give rise to a zone of shear there which may lead to volcanic activity and thus to a volcanic arc.

We may find a certain confirmation of this hypothesis in the distribution of the shallow earthquake foci—*i.e.*, those having regard to the Earth's crust. If the above ideas are right, we may expect these foci to occur in the tectonic zone where the main crustal deformation goes on, as well as in the volcanic belt and in the area between the two where, as it is easy to see, the crust must be subject to strong distortion. Referring to the map of Visser (1930) or to the first map of Koning (1952) giving the centers in the depth zone of 0–50 km (Fig. 3), we see that they are in good agreement with our hypothesis. It is interesting to see that, especially on Visser's map, we find foci north of the arc of the smaller Sunda Islands in the two areas dis-

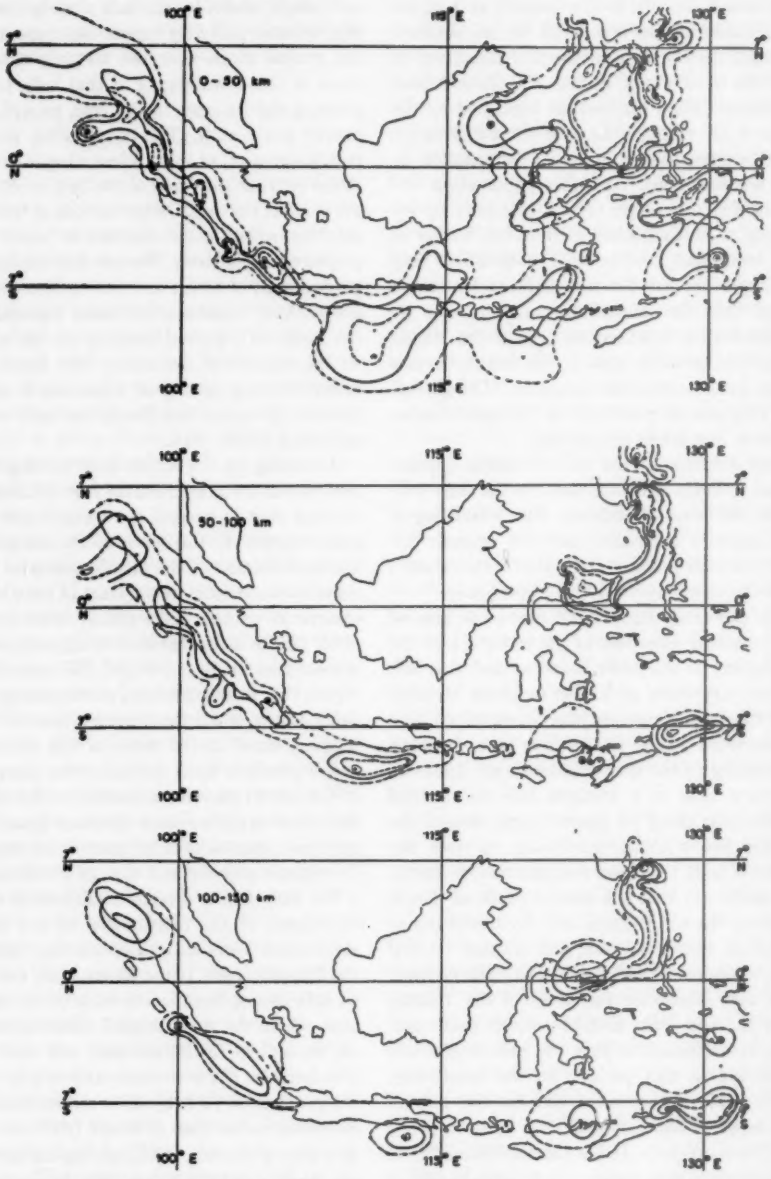


FIGURE 3.—MAPS OF FOCI OF 0–50 KM DEPTH, 50–100 KM DEPTH, AND 100–150 KM DEPTH
(Taken from L. P. G. Koning, Earthquakes, geographical distribution, depths and magnitudes, Proc. K. Ned. Akad. v. Wet. Amsterdam, Ser. B, 55, 1, 1952, pp. 60–77).

cussed above, north of Sumba, and north of Timor; on Koning's map where only the stronger earthquakes are given this is somewhat less clear, and the foci are also less continuous over the whole Banda arc.

Examining likewise Koning's maps for the depth zones 50–100 km and 100–150 km we notice that many foci west of and over Sumatra and south of Java are deeper than 50 km or even than 100 km. This seems to indicate that the deformations caused by the downbulging of the crust—even the slight downbulging west of Sumatra—are continuing deeper in the plastic substratum than might be expected; it is hardly likely that the crustal downbulge south of Java reaches deeper than 50 or 60 km, and west of Sumatra it is probably less.

Returning to our present problem regarding the volcanic arc we may notice that the distance between this arc and the tectonic belt does not vary much; it is usually between 180 and 250 km. We may question why this is the case. The writer should not want to go too deep into this problem here; with our present knowledge it seems difficult to attempt a decisive conclusion. Two possibilities might be mentioned briefly.

(1) This distance might be connected with the deformation of the crust caused by the buoyancy of the negative anomaly belt. It can be shown (Vening Meinesz, 1940) that the upward belts thus brought about on both sides of the belt cause a curvature of the crust involving extra compression in the lower half of the crust and that, therefore, no volcanoes are possible here. They could not thus occur closer to the tectonic belt than where the crustal curvature reverses sign—*i.e.*, on the outside of these adjacent belts which must be marked by positive anomalies; this agrees reasonably well with the distances of 180–250 km.

(2) The location of the volcanic belt may be connected with the shearing stress exerted from below on the crust by currents in the substratum (Vening Meinesz, 1948, p. 54). We shall presently discuss the possibility and even probability of such currents, and it can be shown that the shear on the crust must be expected to attain a maximum at a distance from the tectonic belt of the same order of magnitude. In dealing with these problems the writer will come back to the question here discussed.

It is hardly necessary to point out that the location of the volcanic belt may also be affected by zones of crustal weakness caused by older deformations of the crust; we may expect a tendency to coincide with such zones.

Other Tectonic and Volcanic Belts

We shall not go into the same detail here, as we have hitherto done, with regard to those parts of the tectonic and volcanic belts of Indonesia not yet discussed. The geologic and gravimetric data are less complete elsewhere, and the phenomena are more complicated. We shall only make a few remarks here.

The northern part of the Banda arc ends somewhat unexpectedly near the island of Buru. For understanding this part, a better knowledge of the geology of Ceram and Buru and the gravity fields on both islands would be helpful. We see on the map that the direction of this part of the arc seems to obey well the rule of making an angle of 55° with the sense of the compression, while the part between the Kai islands and Ceram appears to coincide with this sense. The gravity field in the Banda Sea is hardly compatible with the often-expressed idea of a sunken connection from the Tukang Besi islands to Buru; if there is any connection here, as a slight gravity decrease seems to indicate, it rather seems to run toward the volcanic arc, *i.e.*, toward the Banda islands.

As the map shows, the northern arc also seems to agree well with the angle of 55° with the sense of the compression; this is true for the part running west and that running north-northeast from the Molucca Sea. East of the Philippine islands the last tectonic belt changes its direction, enclosing an angle of about 25° with the sense of the compression just as we found for the belt west of Sumatra.

For the northern arc also many data are lacking. A better geological survey and gravity observations in East and Central Celebes would no doubt give us indications how far and with what intensity the deformations of the northern arc continue here. The meager geological data seem to show strong folding on East Celebes,

and so a certain continuation here appears likely. Still the seismic maps of Koning concerning the depth zones of 0–50 km, 50–100 km, and 100–150 km appear to indicate a shear zone running north-northwest or northwest from the end of East Celebes, and this might well point to two separate crustal blocks north of the tectonic belt moving toward it with different velocities and thus indicating that the compressive strain of the belt in the Molucca Sea is larger than that in East Celebes.

According to the first of the three maps of Koning, represented in Figure 3, a second similar zone probably runs from the western end of East Celebes toward the Mangkalihat peninsula of Borneo and from there toward the northwest or north-northwest (hatched line in Pl. 1). It would explain the end of the zone of compression where the East Celebes arm joins the central part of Celebes. Both conclusions can of course not be considered as more than vague indications.

A still more uncertain point is the way the tectonic belt is affected by the Halmaheira crustal block which is probably connected with New Guinea and with the arc running toward the Palau islands and to Yap; we see an interruption between the two patches of strong negative anomalies northwest of the central part of Halmaheira. Here also more data are needed, both on the geology of Halmaheira and Morotai and on the gravity field to the north, east, and southeast. We might then perhaps better understand the fact that there are two volcanic arcs here, one to the northwest over North Celebes and the Sangihe islands and one to the east of the tectonic belt over the islands of Ternate and Batjan, west of Halmaheira. Because of the peculiar triangular shape here of the tectonic belt both arcs seem to be on the concave side of a tectonic curve and so appear to obey the rule mentioned before.

Troughs, Basins, and Third Ridge, Convection-Current Hypothesis

In the whole eastern part of the Indonesian archipelago there seems to be a tendency toward a beginning of the readjustment of the strongly disturbed isostatic equilibrium of the tectonic belt. There are, *e.g.*, clear indications of a strong recent rising of Rotti, Timor, and the islands east of them (Umbgrove, 1949, p. 22), as well as elsewhere. The deep basins on both sides of the belt give the impression of being young features formed by recent sinking. It is obvious to assume these rising and sinking movements to have been caused by the same phenomenon. We may probably conclude that somehow the compression of the crust is less now than it was during the crustal downbuckling, thus allowing the buoyancy of the belt of negative anomalies to come into action by a certain disengagement of the belt from the neighboring crustal zones. These latter zones, originally lifted by the belt's buoyancy, must sink back now and thus form the troughs on both sides of the tectonic belt. We find these deeps on both sides of the whole outer Banda arc and also of the northern arc from East Celebes toward the eastern part of Mindanao.

Besides these linearly shaped troughs on both sides of the tectonic belts we find evidence of recent sinking in the much greater basins inside the arcs, as in the northern and southern parts of the Banda Sea and the Celebes Sea. Most geologists agree in supposing these areas to have been above sea level a few million years ago, while at present they have a depth of about 5000 m. The great depth of nearly 7500 m of the Weber deep between the volcanic and the tectonic arcs east of the Banda Sea may probably be explained by assuming that it came into being by both phenomena combined; this deep thus belongs to the first group of linearly shaped troughs and also forms part of the Banda basin itself.

The problem of the origin of these great deep basins bring us to a new chapter of this paper dealing with the hypothesis of local subcrustal currents. Before starting this chapter, however, we may realize that we have not yet considered the problem of the source of the great field of compression, which we assumed as the cause of the crustal deformations of the Indonesian archipelago. The writer does not want to go deeply into this problem here. He is inclined to account for it by the hypothesis of large-scale convection currents rising under the Asiatic continent because of the higher temperature which can be assumed here in the mantle

of the Earth, and in this area flowing off in the direction of the arrows on the map.

We may suppose that these big current systems are caused mainly by the vertical temperature gradient brought about by the cooling of the Earth and started by temperature differences in a horizontal sense as mentioned. We further must assume that they have made only about half a turn, thus bringing the higher-temperature matter on top and that of lower temperature to the bottom. In 1939 Griggs pointed out that we may thus explain the greater and smaller rhythms of the tectonic activity of the Earth's crust; after the stability has been re-established by a current system a long time ensues before the cooling again brings about instability and before a horizontal gradient acts as trigger.

By admitting these currents in the mantle we can also explain the regularities which appear to exist in the great lines of the Earth's topography—*i.e.*, in the distribution of the continents and oceans (Vening Meinesz, 1951a; 1952a; 1952b). This is in harmony with a distribution of the currents as corresponding to the thickness of the mantle of about 2900 km, supposing these currents to sweep together the sialic crust in blocks forming the continents. It would lead us too far outside our subject to elaborate on these ideas here.

There are strong indications for smaller subcrustal currents likewise. In the first place we have the deep and intermediate earthquake foci proving the existence of strong shear stresses in the Earth up to depths of more than 700 km. Examining the maps of Koning (1952) we see deep foci in belts from the Java Sea to the Banda Sea and in the Celebes Sea and West Mindanao. We thus find the rule confirmed that everywhere they occur on the continental side of the tectonic belt. For three of these deep-focus earthquakes the mechanism in the focus has been determined (Koning, 1942; Ritsema, 1952), and the results agree well. They all pointed to shear along planes dipping under the continental side at an angle of 55° or in planes intersecting these planes probably along a horizontal line and at right angles. North of Lombok the strike of these planes was N. 60°E., south of the Tukang Besi islands N. 50°E., and in West Mindanao N. 20°W. In all three cases the movement corresponded to horizontal tension and vertical compression.

These results are contradictory to the view often expressed that the deep earthquakes are caused by a push of the ocean crust under the continent along a shear plane dipping under the continent; the relative movement has the opposite direction. The distribution of the deep foci in the vertical profiles of Koning at right angles to the tectonic belt is likewise not in agreement with this plane; the deep foci are found in a more or less rectangular patch, and the shallow and intermediate foci usually form only one patch of a more irregular shape, which last fact we can readily understand in view of the varying properties of the different crustal and subcrustal layers in action here.

The indication of vertical compression in a deep layer between a depth of 300–400 km and one of 500–700 km and situated below the recently sunken basin seems to point to a downward current which may likewise be responsible for the sinking. On the other hand there has been recent rising, though by a smaller amount, in the arcs surrounding the basins, and so we get the impression of a rising current here; most intermediate foci found near the volcanic or tectonic arcs seem to confirm this view. The data about these earthquakes, however, were not sufficient for an investigation about the mechanism in their focus, and so it is not possible to say whether this is also in harmony with our hypothesis.

However, still another phenomenon seems to confirm it. We have seen that the rising of the tectonic belt combined with the sinking of the trough-shaped basins on both sides can probably be interpreted as a beginning of the readjustment of the isostatic balance, and that this could be the case only if the compression in the crust was somehow less than during the period of strong crustal deformation in the tectonic belt. This is in harmony with the supposition of a rising subcrustal current below this belt because such a current must be expected to flow off horizontally below the crust and to cause tension there. The "graben" developed during the Pliocene on Timor and other islands of the tectonic belt might well be interpreted as caused by the beginning of this tension effect.

So we see there is converging evidence of

recent current systems in the substratum that sink below the basins and rise under the surrounding arcs. We might perhaps find a cause of such currents in the Earth's cooling coupled with an increase of temperature below the

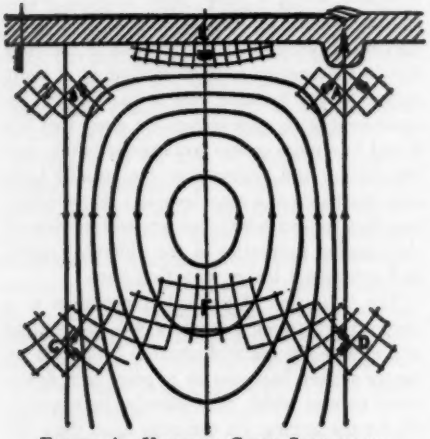


FIGURE 4.—VERTICAL CROSS SECTION OF CONVECTION CURRENT

Showing areas of maximum shear, *i.e.*, of drag exerted on the crust and of deep and intermediate foci.

After Meinesz (1948)

tectonic belt brought about by the concentration there, because of the pushing together of the crust, of sialic matter richer in radioactive constituents than the substratum. An investigation based on the hydrodynamical equations for viscous fluids led to a current pattern (Vening Meinesz, 1948, p. 45 *et. seq.*) as given by Figure 4.

The question arises whether such currents could also arise when the tectonic belt is situated in an area where the crust consists nearly entirely of ultrabasic rocks and where, therefore, the crustal downbuckling gives a deep trough at the surface. In that case the trigger effect caused by the difference of temperature in a horizontal sense brought about by the concentration of sial in the tectonic belt would perhaps be missing or, if there would also be in this case some concentration of radioactivity, it would probably be much slower to come into action. We shall not pursue this problem here; it is not present in Indonesia.

We must now examine the question whether the seismological data can be reconciled to the hypothesis of convection in a subcrustal layer many hundreds of kilometers thick. As is well known, recent seismological studies (Birch, 1951) have revealed that the densities in the mantle of the Earth, after reducing them to surface conditions of pressure, show possibly a constant value of about 3.3 in a zone to a depth of 200 km and, still more likely, a constant one of 4.0 from about 950 km down to the core, but a gradual transition between those values in the area between 200 and 950 km depth. The question, therefore, is whether a current would be able to break through this zone of density change.

If the transition is a change of phase by a rearrangement of the atoms, supposed to be reversible, we may assume the current could break through. It is likely that such a transition occurs at great speed when the temperature reaches a critical value dependent on the pressure, and we shall suppose this value increases with increasing pressure and, therefore, with increasing depth z; we shall indicate this temperature function of z by t_z. We shall further assume the transition from the lighter phase to the denser one to be accompanied by a liberation of energy, and the reverse to be true for the transition from denser to less dense; these amounts of energy must increase considerably with pressure and, therefore, with depth.

We may suppose that the mantle consists of olivine or a similar type of rock in a rhombic phase which changes with increasing pressure in a denser one, perhaps to a cubic phase, but this hypothesis does not affect our problem as long as the above conditions are fulfilled.

In that case we may assume the temperature in the transition zone to be given by t_z and the ratio of both phases at each depth to be determined by what has taken place in the past. If we suppose this part of the mantle to be slowly cooling and the corresponding outward heat flow to diminish downward, we must conclude that the necessary excess of heat for each particle needed to keep this flow going is provided by the transition of a small part of the particle into the denser phase. We must, therefore, suppose that there is an extremely slow increase of the denser phase in the transition

zone and that this zone is thus slowly migrating outward. As no figures are known for the physical quantities implied in this process, it is hardly possible to estimate its speed, but there can be no doubt about the extreme slowness of the phenomenon.

Our hypothesis has some remarkable implications for the supposition of subcrustal currents. Assuming a trigger effect by the slow heating of the substratum in the neighborhood of the tectonic belt as mentioned above, we can explain the long time lag of 15–20 million years in the basin development with regard to the last period of strong folding in the upper Miocene, assumed to be responsible for the downbulging of the crust (Vening Meinesz, 1947b; 1948, p. 41 *et. seq.*).

We can further conclude that this heating must bring about a swelling of the substratum near the tectonic belt and the lateral flowing off of part of this swelling, which we can well understand to occur toward the inside of the arc where the resistance must be less than toward the outside. The resulting load on the subcrustal layer inside the arc may be expected to result in a sinking of this layer, and, because of the downward temperature gradient, this must cause a lowering of the temperature at a certain fixed depth z in the whole area of the sinking column. The result must be that in the transition layer a corresponding change toward the denser phase takes place until the liberation of energy has restored the temperature t_z; this must be accompanied by a sinking of the surface. We thus can account for the origin of the deep basin.

On the other hand the subcrustal column under the tectonic belt has been unloaded by the flowing off of matter at the surface and must have a tendency to rise. The resulting increase of temperature must bring about the reverse process—*viz.*, a change into the lighter phase and a rise of the surface—, and this must lead to an acceleration of the subcrustal flow toward the basin. Deeper down the loading of the basin column must cause an excess of pressure and the unloading of the column under the tectonic arc a decrease of pressure, which must bring about a current from the first area to the last, closing the current circuit.

We thus see a sort of heat-transport current coming into being, similar to a convection current and breaking through the transition layer. Interesting conclusions are possible. In the first place we may suppose the change of matter toward the denser phase in the sinking column to be final; the sinking has played the part of accelerating the slow natural cooling. The deep basin at the surface is, therefore, likely to continue to exist, and this is in harmony with the fact that the geologists find many such basins of old origin, as, *e.g.*, the basin of Paris. The view formerly held by the writer that the basins have been caused by normal half-turn convection does not allow such a simple explanation of their permanent character.

According to our hypothesis such a permanent character can not be ascribed to the results brought about by the heating in the rising column. The change to the lighter phase must have been impeded by the cooling during the existence of the current, and this effect must also have continued to reduce the amount of lighter matter. Probably we may, at least partly, attribute to this cause the fact that the arc has risen less than the basin has been sinking.

We have already briefly discussed that our hypothesis also explains the deep and intermediate earthquake foci and the mechanisms found in the foci of three of these deep earthquakes. It fits even somewhat better than that of a normal half-turn convection current as it makes it clearer that the foci do not, in general, occur in the two other angles of the more or less rectangular current pattern (Fig. 4; Vening Meinesz, 1948, p. 51 *et. seq.*).

In a future paper the writer hopes to cover in more detail these problems and consequences of the hypothesis of heat-transport currents in connection with a two-phase character of the Earth's mantle and also discuss more thoroughly the different maps and profiles of Koning's (1952) study of earthquake foci. Here he wants only to mention two more subjects which can also be explained by the subcrustal-current supposition.

A third ridge occurs on the inside of the volcanic arc in many island-arc areas, as, *e.g.*, the Aves ridge in the Caribbean area, the ridge west of the Marianas arc, and the Turtle and Lucipara islands ridge in the Banda Sea (Vening Meinesz, 1951b). Probably this

last ridge continues toward the Tukang Besi islands and the islands southeast, south, and southwest of the Tiger islands and follows the northern edge of the Flores Deep and the Bali Sea, over the Postiljon islands and the Paternoster islands. Possibly we may even recognize it farther westward in the island of Madura and over the northern border of the oil-bearing idiogeosyncline of northern Java. The ridge is parallel to the outer tectonic belt and follows rather closely its two great curves near Timor and near Sumba. As has been discussed in the above-mentioned paper, the development of this ridge might be attributed to a pushing together of the crust by the strong shear stresses exerted on it by the subcrustal current here dealt with; if we try to determine the compression in the crust brought about by these shear stresses we find stresses of the order of 10,000 kg/cm², and it is improbable that the crust can stand this. That it does not bring about a complete tectonic geosyncline development can be understood if we realize that the current here discussed cannot last longer than the time needed for a half turn and that this does not take more than a few million years. For development of a complete geosyncline we need forces lasting at least a few tens of millions of years, and so we must then suppose current systems of much larger dimensions reaching down over a great part if not the whole thickness of the mantle. We have already mentioned the possibility that such currents have occurred.

From our supposition it follows that we may expect the third ridge to have a tendency to develop more or less above the sinking column of the current—*i.e.*, in the area where the deep earthquake foci are found.

We may lastly point out that with the two types of basins hitherto dealt with we have not covered all the young basin formations of the eastern part of the Indonesian archipelago. We have not yet discussed the basins of Strait Makassar between Borneo and Celebes, of the Gulf of Bone between South and Southeast Celebes, and of the Gulf of Tomini between North and East Celebes. It is remarkable that all these basins have a depth of about 2000 to 2500 m. As Kuenen (1934) has already remarked in his notable discussion of the basins of the archipelago in which he comes to the same grouping of basins as here advocated on geophysical grounds, the latter basins form a group comparable to that of the large deep basins of the Celebes Sea and the two parts of the Banda Sea; though about half as deep they have the same character of having a fairly flat bottom and relatively steep sides.

This similarity points possibly to the same origin and so, if this is indeed the case, we might perhaps suppose them to have likewise been caused by subcrustal currents and, more specifically by the sinking column of these currents. The theoretical treatment of the current equations shows that we must expect a certain ratio between the breadth and the depth of these currents, and as the last type of basins is slightly less than half as deep as the preceding type of basins this is in harmony with the fact that the ratio of the breadths of both groups is roughly the same. This leads to the suggestion that the depths of the currents would likewise have this same ratio. For our last type of basins this would point to a depth of the subcrustal currents of about 200–250 km. It is remarkable that this is just the thickness of the subcrustal layer of homogeneous density of about 3.3 under surface conditions.

The question of course arises about the origin of such currents. We may perhaps again suppose them to have their main cause in the cooling of the Earth, but, as the corresponding basins are not enclosed by tectonic arcs, as is the case for the deep basins, we cannot assume the same trigger effect which in the last case was supposed to be brought about by the heat produced by the concentration of sialic matter in the tectonic belts. We see, however, that all these basins are branches of the deeper basins; the Strait of Makassar is connected in the north to the Celebes Sea and in the south to the Flores basin which itself may be seen as the continuation to the west of the Banda basin, the Gulf of Bone is a branch of the connection between the two, and the Gulf of Tomini of the Molucca Sea. This leads us to the hypothesis that the current movement below the big basins has propagated itself by viscous friction in these directions and that this constitutes the necessary trigger effect for the currents in the homogeneous subcrustal layer. These secondary cur-

rent systems must have caused temperature deviations in the substratum and thus have brought about the sinking of the smaller basins and perhaps also some rising of the adjacent areas.

Propagation of the current system as here supposed may perhaps also have played a part in the spreading of the bigger current systems assumed under the deeper and larger basins. The continuation of the Banda basin toward the narrower Flores deep, *e.g.*, might perhaps have been promoted in this way.

According to the views here given about the two groups of flat-bottomed basins in the Indonesian Archipelago, the larger ones of a depth of about 5000 m and the smaller ones of a depth of about 2000 m, we may conclude that the remarkable constancy and equality of their depths might be attributed, for the smaller basins, to the constant thickness in this area of the subcrustal layer of homogeneous density and, for the larger basins, to the same conditions of the transition zone below this subcrustal layer which must have determined the depths to which the currents reached and the vertical dimensions of the resulting density changes leading to the sinking of the basin floor.

Idiogeosynclines

We have nearly finished our short geophysical survey of the problems regarding the more recent formations in the Indonesian archipelago. There is only one feature not yet mentioned which the writer wishes to touch on—the development of the idiogeosynclines in North Java, the probably older one of East Sumatra (Miocene), and the still older ones of East Borneo and North Sumatra (Eocene) (Umbgrove 1949, p. 44). We might perhaps attribute them to the development of a second crustal wave inside the downbuckling wave of the tectonic arc.

Insofar as the Java idiogeosyncline is concerned, it strikes us, however, that we find deep earthquake foci north of that island pointing to subcrustal currents as dealt with here. This might indicate that we can perhaps consider the Java idiogeosyncline as caused in the same way as the Banda basin of which it is the continuation to the west via the Flores deep. This explanation would coincide with the view taken recently by Umbgrove (1948). In this case the adjacent ridge in North Java may be interpreted as the continuation toward the west of the third ridge. It certainly is the same distance from the tectonic belt as the third ridge of the Turtle and Lucipara Islands in the Banda Sea.

This brings our geophysical survey of Indonesia to a close. The writer feels conscious of the hypothetical character of many views here advanced. Still it may have its use to put them forward, be it only for stimulating discussions and, more important still, for promoting research.

References Cited

Anderson, E. M., 1942, Dynamics of faulting and dyke formation: Edinburgh and London, Oliver and Boyd, 199 p.

Birch, Francis, 1951, Remarks on the structure of the Mantle, and its bearing upon the possibility of convection currents: Am. Geophys. Union Trans, p. 533.

Bylaard, P. P., 1936, Théorie des déformations plastiques et locales par rapport aux anomalies négatives de la gravitation, etc.: Report to Intern. Union of Geodesy and Geophysics, Intern. Assoc. of Geodesy, Congress Edinburgh.

Griggs, David, 1939, A theory of mountain building: Am. Jour. Sci., v. 237, 611–650.

Hess, Harry H., 1938, Gravity anomalies and island arc structure with particular reference to the West Indies: Am. Philos. Soc. Proc., v. 79, no. 1, p. 71–96.

—— 1939, Recent advances in interpretation of gravity anomalies and island arc structure: Adv. Rept. Comm. on Continental and Oceanic structure I. U. G. G., Washington.

Hubbert, M. King, 1951, Mechanical basis for certain familiar geologic structures, Geol. Soc. America Bull., v. 62, p. 355–372.

Kennedy, W. Q., and Anderson, E. M., 1938, Crustal layers and the origin of magmas: Bull. Volcanologique, Série II, III, p. 24–82.

Koning, L. P. G., 1942, On the determination of the fault-planes in the hypocenter of the deepfocus earthquake of June 29, 1934 in the Neth. E. Indies: K. Ned. Akad. v. Wet. Amsterdam Proc., v. 45, no. 7, p. 636–642.

—— 1952, Earthquakes in relation to their geographical distribution, depth and magnitude, I, the East Indian Archipelago: K. Ned. Akad. v. Wet. Amsterdam Proc., Ser. B., v. 55, no. 1, p. 60–77.

Kuenen, Ph. H., 1934, Gravity expeditions at sea, Vol. II, Relations between Submarine Topography and Gravity Field: Pub. Neth. Geod. Comm., Delft, p. 183–193.

Muller, J. J. A., 1895, De verplaatsing van enige triangulatiepilaren tengevolge van de aardbeving van 1–7 Mei 1892: Verh. Kon. Akad. v. Wetensch. Amsterdam, 1 Sectie, III 2.

Ritsema, A. R., 1952, Over diepe Aardbevingen in de Indische Archipel: Dr. thesis, Univ. of Utrecht.

de Sitter, L. U., 1952, Pliocene uplift of Tertiary mountain chains: Am. Jour. Sci., v. 250, p. 297–307.

Umbgrove, J. H. F., 1938, Geological history of the East Indies: Am. Assoc. Petroleum Geologists Bull., v. 22, p. 1–70.

—— 1948, The origin of deep-sea troughs in the East Indies: 18th Sess. Intern. Geol. Cong. Rept., London, pt. VIII.

Umbgrove, J. H. F., 1949, Structural history of the East Indies: Univ. Press, Cambridge, England.

Van der Vlerk, I. M., and Umbgrove, J. H. F., 1927, Tertiaire gidsforaminiferen: Wetensch. Meded. no. 6.

Van der Vlerk, I. M. and Leupold, W., 1931, "Tertiary", Martin-Feestbundel: Leidsche Geol. Meded. Dl. V.

Vening Meinesz, F. A., 1940, The Earth's crust deformation in the East Indies: K. Ned. Akad. v. Wet. Amsterdam, Proc. v. 43, no. 3, p. 278–293.

—— 1941, Tables for regional and local isostatic reduction (Airy system): Pub. Neth. Geod. Comm., Delft.

—— 1943, Spanningen in de aardkorst tengevolge van poolverschuivingen: K. Ned. Akad. v. Wet. Amsterdam, v. 52, no. 5, p. 185–196.

—— 1947a, Shear patterns of the Earth's crust: Am. Geophys. Union Trans., vol. 28, no. 1, p. 1–61.

—— 1947b, Convection-currents in the Earth: K. Ned. Akad. v. Wet. Amsterdam Proc., ser. B. v. 50, no. 3, p. 237–245.

—— 1948, Gravity Expeditions at Sea, Vol. IV: Pub. Neth. Geod. Comm., Delft.

—— 1950, Earth's crust deformations in geosynclines, K. Ned. Akad. v. Wet. Amsterdam Proc., ser. B, v. 53, no. 1, p. 27–46.

—— 1951a, A remarkable feature of the Earth's topography, origin of continents and oceans; K. Ned. Akad. v. Wet. Amsterdam Proc., ser. B, v. 54, no. 3, p. 212–228.

—— 1951b, A third arc in many island arc areas: K. Ned. Akad. v. Wet. Amsterdam Proc., ser. B, v. 54, no. 5, p. 432–442.

—— 1952a, The origin of continents and oceans: Geologie en Mijnbouw, n. ser., v. 14, no. 11, p. 373–384.

—— 1952b, Convection currents in the Earth and the origin of the continents: K. Ned. Akad. v. Wet. Amsterdam Proc., ser. B, v. 55, no. 5, p. 527–553.

Vening Meinesz, F. A., Umbgrove, J. H. F., Kuenen, Ph. H., 1934, Gravity expeditions at sea, Vol. II: Pub. Neth. Geod. Comm., Delft.

Visser, S. W., 1930, On the distribution of earthquakes in the Netherlands East Indian Archipelago: Verh. Kon. Magnet. en Meteorolog. Observat. Batavia, v. 22.

Potgieterlaan 5, Amersfoort, Nederland

Manuscript Received by the Secretary of the Society, June 2, 1952

Proc.,

ol. IV:

in geo-
sterdam

Earth's
oceans;
oc., ser.

c areas:
oc., ser.

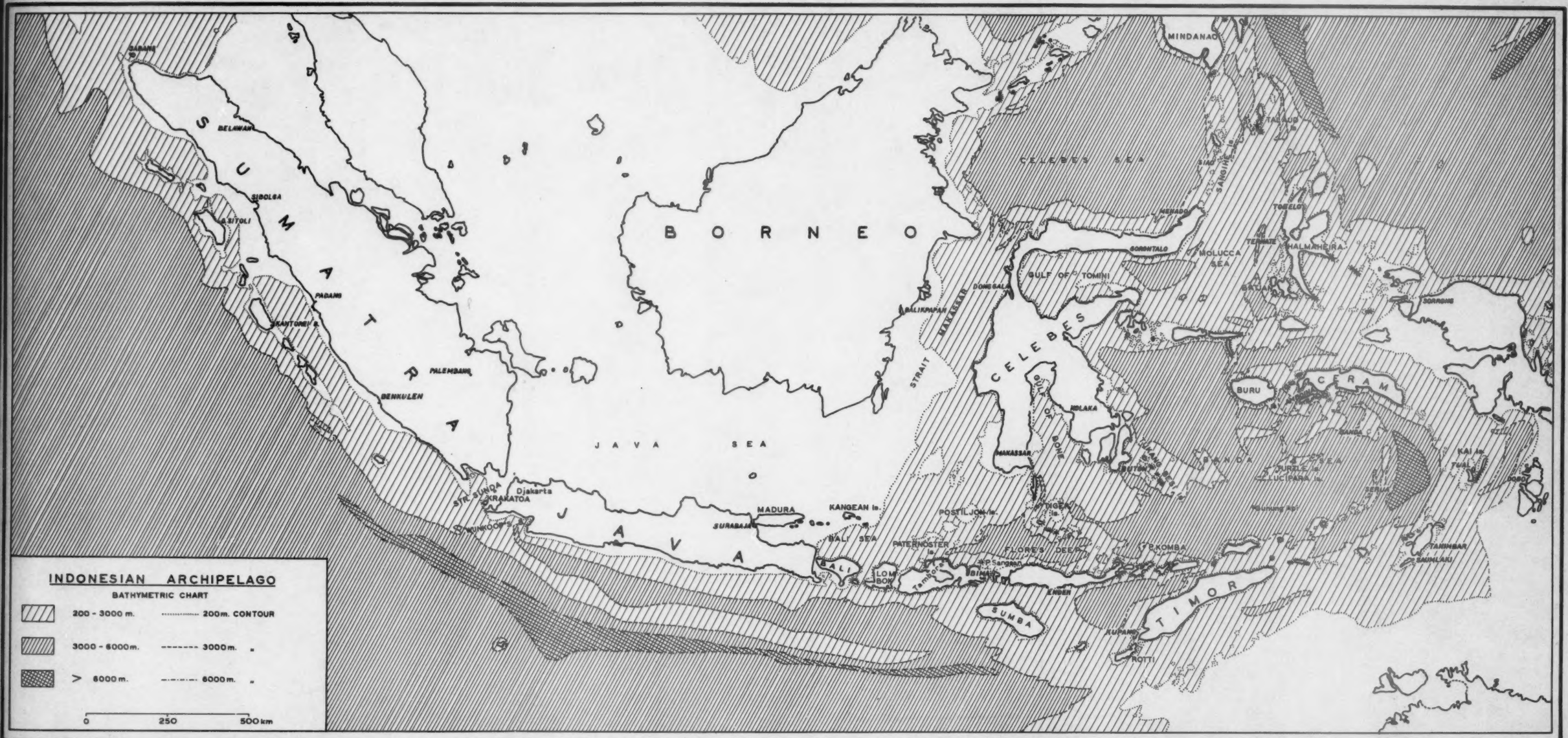
oceans:
o. 11, p.

rth and
Akad. v.
, no. 5,

Kuenen,
sea, Vol.

of earth-
n Archi-
teorolog.

ND
Y OF THE



CONTOUR MAP SHOWING BASINS OF INDONESIAN ARCHIPELAGO

S
ocea
disc
leve
sligh
thic
the
T
that
gran
lava
juve
debr
form

Intr
Ack
Seis
Turi
Gra
Stru
d
Hyp
sc
Imp
Ref

N
the
min
isla
(19
pec
T
larg
exp
vat
pre
ava
cat
bas

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 65, PP. 165-174, 2 FIGS. FEBRUARY 1954

GRAVITY ANOMALIES AND STRUCTURE OF THE WEST INDIES PART I

BY MAURICE EWING AND J. LAMAR WORZEL

ABSTRACT

Seismic-refraction results and gravity data have been used to deduce the crustal structure from the ocean basin north of the Puerto Rico trench to the Caribbean Sea. It is concluded that the Mohorovičić discontinuity (characterized by compressional-wave velocities of about 8 km/sec) lies at 9 km. below sea level under the ocean basin, 12 km under the Caribbean Sea, at about 16 km under the trench, and at slightly shallower depth under Puerto Rico. The large negative gravity anomaly is attributed to a great thickness of sediments in the trench rather than to a "sialic root" due to a down-buckle of the crust under the trench, as formerly thought.

Turbidity currents are assigned an important role in the accumulation of the sediments. It is suggested that a trench formed in an unspecified way quickly collects sediments, largely by turbidity currents. When granitized and uplifted the sediments form an island arc like the West Indies. Contamination of basaltic lavas by the sediments can account for andesitic lavas, and the accompanying water, rather than being juvenile, is derived from sea water. The trenches at or near the continental margins confine continental debris to the continental margins and collect oceanic debris. The basaltic crust and this debris are first formed into an island arc and later into a continental addition.

CONTENTS

TEXT

	Page
Introduction	165
Acknowledgments	165
Seismic results	166
Turbidity currents	169
Gravity results	169
Structure deduced from seismic and gravity data	170
Hypothesis on the origin of the island arc and some consequences of it	171
Implications for continental growth	172
References	172

ILLUSTRATIONS

Figure	Page
1.—Chart of Puerto Rico area	166
2.—Structure and gravity sections true north of San Juan, Puerto Rico	167

TABLE

Table	Page
1.—Isostatic anomalies in the Puerto Rico area	170

INTRODUCTION

New seismic and coring evidence indicates the need for a reconsideration of the gravity minimum associated with the West Indian island arc. Ewing (1952) and Vening Meinesz (1954) have already written on some of the aspects of the new data.

The seismic data are far from complete, owing largely to the failure to receive a shipment of explosives. It is expected that additional observations will be made in the next year. The present preliminary paper is offered to make available in written form the verbal communication on which Vening Meinesz (1954) has based important parts of his paper.

ACKNOWLEDGMENTS

Mr. George Sutton generously made available in advance of publication most of the seismic results used here. Mr. Bruce Heezen gathered together all the sounding data, did the contouring, and very kindly made his chart available in advance of publishing the topography of the Puerto Rico trench.

We wish to acknowledge the contributions of sounding data and the earliest seismic work in the area by Dr. J. B. Hersey.

Most of the seismic work was done under Contract N6-onr-271 with the Office of Naval Research and NObsr 43355 with the Bureau of Ships, both of the U. S. Navy.

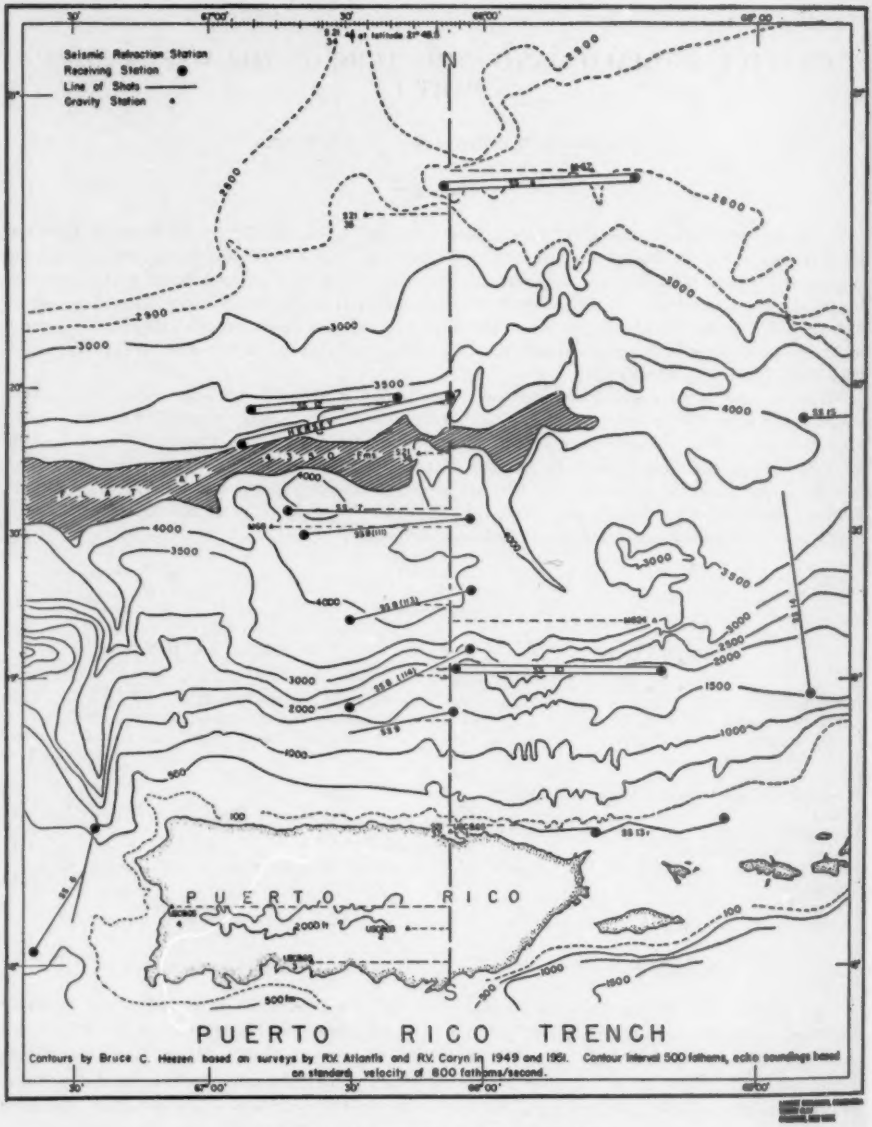

Seismic Refraction Station
Receiving Station
Line of Shots
Gravity Station
N
PUERTO RICO
PUERTO RICO TRENCH
Contours by Bruce C. Heezen based on surveys by RV. Atlantis and RV. Coryn in 1949 and 1951. Contour interval 500 fathoms, echo soundings based on standard velocity of 800 fathoms/second.

FIGURE 1.—CHART OF PUERTO RICO AREA

SEISMIC RESULTS

Six seismic-refraction measurements were made in the spring of 1951 in and near the trench north of Puerto Rico. These together with a measurement made by Hersey in 1949 are shown in Figure 1 superimposed on the best bathymetric chart available for publication, H.O. 5487. The locations of the gravity stations observed by Vening Meinesz and the U. S. Coast and Geodetic Survey are also shown. A

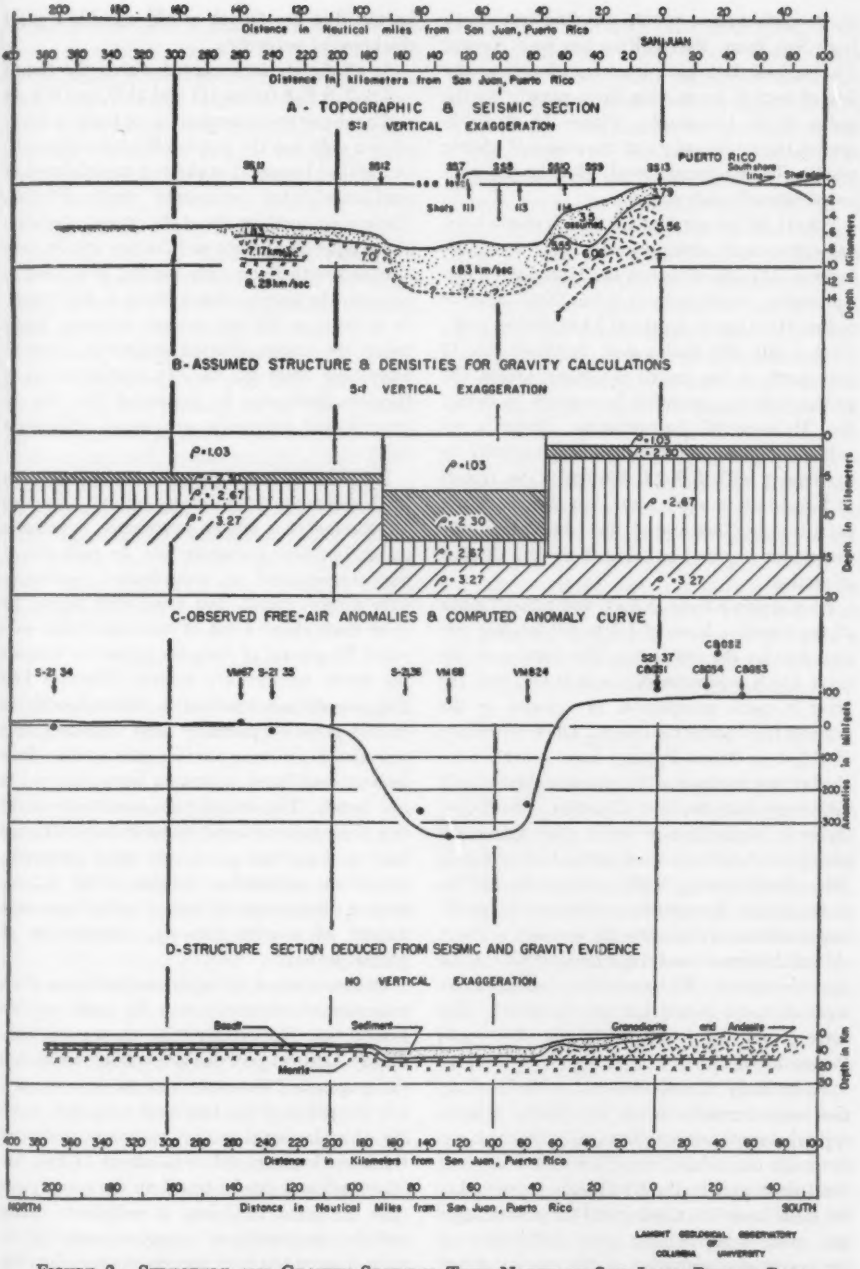

Distance in Nautical miles from San Juan, Puerto Rico
SAN JUAN
Distance in kilometers from San Juan, Puerto Rico
A - TOPOGRAPHIC & SEISMIC SECTION
5:1 VERTICAL EXAGGERATION
PUERTO RICO
Depth in Kilometers
B-ASSUMED STRUCTURE & DENSITIES FOR GRAVITY CALCULATIONS
5:1 VERTICAL EXAGGERATION
C-OBSERVED FREE-AIR ANOMALIES & COMPUTED ANOMALY CURVE
Anomalies in milligals
D-STRUCTURE SECTION DEDUCED FROM SEISMIC AND GRAVITY EVIDENCE
NO VERTICAL EXAGGERATION
Basalt
Sediment
Granodiorite and Andesite
Mantle
Depth in Km
NORTH
Distance in Nautical Miles from San Juan, Puerto Rico
SOUTH
LAMONT GEOLOGICAL OBSERVATORY OF COLUMBIA UNIVERSITY

FIGURE 2.—STRUCTURE AND GRAVITY SECTIONS TRUE NORTH OF SAN JUAN, PUERTO RICO

north-south line of profile trending true north from San Juan, Puerto Rico has been chosen. The various data have been transferred to the line of section by moving them parallel to the strike of the topography. Figure 2A shows in section the topography and the observed seismic results. The observed sound velocities are indicated beneath each station.

S.S.-11 80 km north of the trench shows typical ocean-basin structure—*i.e.*, a 3 km-thick layer of 7.17 km/sec above the Mohorovičić discontinuity, overlain by a 1 km-thick layer of sediments in ocean depths of 5 km (Officer *et al.*, 1952, p. 801, 802; Ewing *et al.*, In press). S.S.-12 just north of the trench is similar, except the profile was not extended far enough to detect the Mohorovičić discontinuity. Hersey's results are, though fragmentary, essentially in agreement with S.S.-12. Similarly the results at S.S.-15 are in agreement, but they have not been included because of the large lateral displacement required to bring them into the line of section.

At stations S.S.-13, S.S.-9, and S.S.-10 south of the trench a layer of 5.5 to 6.0 km/sec was found under the sediments. The bottom of this layer has a *minimum* depth of 9 km, and the layer is quite comparable in velocity to the granitic layer under continents. The seismic data of S.S.-6 in Mona Passage have not been included here because of the obvious topographic difference from the line of section. The deeper layers in Mona Passage are in good agreement with S.S.-13. Seismic work in the Virgin Islands (Worzel and Ewing, 1948, p. 46) about 150 km to the east of the present section shows consolidated sediments (velocity 4.0 km/sec) of about .45 km thickness overlying a layer of 5.6 km/sec near the islands. The consolidated sedimentary layer thickens greatly toward the trench. This corresponds very well with the data near Puerto Rico.

Preliminary calculations on seismic-refraction measurements in the Caribbean indicate typical ocean-basin structure, as outlined above. Since the ocean-basin structure found north of the trench and in the Caribbean showed that the sialic layer was missing and the simatic layer was quite thin, making great difficulties for the accepted explanation of the gravity deficit in terms of a crustal down-buckle, it was expected that the trench would contain a great thickness of sediments.

At all the stations located within the trench —S.S.-7, S.S.-8 (shots 111 and 113), and S.S.-14 —the seismic measurements gave positive information only for the unconsolidated sediments; i.e. neither basement rocks nor consolidated or semi-consolidated sediments were detected. There is no question about the proper operation of the equipment since satisfactory results were obtained with it on days before, after, and in between the several observations in the trench. In more than 100 sea seismic stations, layers below the unconsolidated sediments have always been observed. On any interpretation of these results it must be concluded that the unconsolidated sediments are many kilometers thick.

Inability to map basement rocks beneath a thick sedimentary section by seismic-refraction measurements is not an uncommon experience on land or near the shore line. In such places, semi-consolidated or consolidated sediments were always found, and none were buried by more than about 1 km of unconsolidated sediment. In general at deep-sea refraction stations the entire sedimentary column (about 1 km) is apparently unconsolidated, although evidence on this point is inherently hard to obtain. It is remarkable that no velocities greater than those for unconsolidated sediments were observed in the trench. The compaction commonly found in a thick sedimentary column has not occurred here, perhaps because of very rapid accumulation of the sediment or because of the dependence of compaction on factors other than mere weight of overburden—*e.g.*, percolation of ground water.

One estimate of the minimum thickness of unconsolidated sediments can be made on the assumption that the explosive charge was sufficiently large to produce a readable signal, but the longer-shot distances were too short to permit reception of the basement-refracted wave. On this assumption the minimum sediment thickness is computed to be about 12 km. Another estimate can be based on the assumption that the great thickness of sediments causes sufficient absorption of energy to make the refracted wave too weak for observation. In effect this amounts to declaring that the longer shots

of each profile must be excluded from the calculation of minimum depth. For many reasons it is impossible to make precise calculations of the absorption. A rough estimate may be made as follows: For comparable conditions at a typical deep-sea station the longest horizontal travel for the refracted wave travelling in unconsolidated sediments is 8–16 km. Assuming vertical and horizontal absorption are approximately equal, and that no other factors enter, the minimum thickness of sediments in the trench would be 4–8 km. Additional seismic-refraction measurements in the trench are needed.

Barbados, which lies on a continuation of the gravity minimum associated with the Puerto Rico trench, was investigated in 1944 (Worzel and Ewing, 1948, p. 48), and the basement rocks were not detected. It was concluded that about 1 km of unconsolidated sediments and 1 km of semiconsolidated sediments covered more than 3 km of consolidated sediments. This is consistent with the idea that the sediments were laid down under conditions now prevailing in the Puerto Rico trench.

Turbidity Currents

Ericson *et al.* (1951; 1952) and Heezen and Ewing (1952) have presented abundant evidence of the widespread operation, in the late Cenozoic oceans, of turbidity currents capable of carrying sediments hundreds of miles. The types of evidence of turbidity currents presented are sands containing shallow-water species of Foraminifera interbedded with oozes of abyssal facies, graded bedding in the turbidity-deposited layers with sharp lower boundaries, high-carbonate gray clays repeatedly alternating with red clays in basins, particles of shallow-water flora and fauna of Pleistocene and Recent age found in the turbidity-deposited layers at depths greater than 2000 fathoms, mineral and faunal similarity of deep-sea sands and continental-shelf sediments, and finally the flat floors of the deepest depressions being formed of numerous sedimentary layers exhibiting many alternations of colors and sediments.

The two cores (±10 feet long) taken in the flat plain on the bottom of the Puerto Rico trench clearly show deposition of much of the sediment by turbidity currents (Ericson *et al.*, 1952, p. 496–498, 509). These cores consist largely of thick graded calcareous sands containing shallow-water fauna and flora indicating a very large contribution of sediment to this trough from the surrounding shallow-water areas. Red clays are interbedded with the calcareous sands. Three cores on the north wall of the trench are normal deep-sea red clay throughout. Two cores on the south wall are essentially all clay, showing evidence of slumping. One is Pleistocene, the other Miocene. All of these cores are from depths so great that they would be expected to be red clay throughout if the deposition occurred in the absence of transportation along the bottom. Available bathymetric data indicate that a continuous ridge caps the north and east walls of the trench, cutting it off from turbidity currents from the main ocean basin. It is improbable that the barrier has been effective throughout the existence of the trench. Hence, if one grants the possibility of turbidity currents from the main ocean basin to the trench, it is reasonable to suppose that much of the sedimentary column contains a much smaller percentage of calcium carbonate than that reached by the coring tubes.

Gravity Results

Reconsideration of Vening Meinesz' (1934, p. 117–133; 1948, p. 25–90) and Hess's (1933; 1938, p. 71–86; 1951) hypothesis of the sialic crust down-buckling to form a local thickening of the sialic crust (tectogene) is required because of the seismic-refraction evidence that the sialic crust is absent under the Atlantic Ocean and under the Caribbean Sea. Vening Meinesz required a cross section of 1200 km² for a tectogene with a density contrast of 0.6 at the base of a 25- or 30-km crust, to produce the observed anomaly, corresponding to a crustal shortening of 50 or 40 km respectively. If the 3- to 5-km-thick layer found by seismic-refraction methods to overlie the Mohorovičić discontinuity and to underlie the sediments of the ocean floor was the source of the thickened section, the shortening required would be 250 to 500 km. The 1200 km² block calculated by Vening Meinesz was for the Java-Christmas Island profile where the largest isostatic anomaly was −130 mgals,

whereas the largest isostatic anomaly reported by Vening Meinesz (1948, p. 230) for the trench north of Puerto Rico was −226 mgals (the largest negative anomaly reported to date). Thus more than 500 km of shortening would be required to produce the anomaly. Hess (1933, p. 31) prefers a crust with a density 0.3 less than the material beneath. This figure doubles the required crustal shortening.

Table 1.—Isostatic Anomalies in the Puerto Rico Area

Station		Water depth (Meters)	Free-air anomaly (Mgals)	Hayford isostatic anomaly (Mgals)	Airy local isostatic anomaly (T = 30 km)
S-21	34	5569	−18	−1	
	35	5340	+21	+31	
	36	7699	−264	−125	
	37	9	+129	−19	
Vening Meinesz	67	5510	+9	+7	+7
	68	8040	−341	−193	−292
	824	7370	−244	−226	−222
Elevations					
U. S. Coast & Geodetic Survey	1	+8	+112	−43	On land
	2	+648	+220	+23	
	3	+14	+91	−48	
	4	+16	+122	−20	

The location of the gravity stations in the Puerto Rico area is shown in Figure 1. Vening Meinesz first discovered the negative gravity trough associated with this topographic trench on his world-circling cruise of Hr. Ms. K-XIII. On the cruise of the S-21 he made additional observations and again on Hr. Ms. 0 13. Heiskanen (1939, p. 98) has given the best values for the S-21 data, while Vening Meinesz (1948, p. 156, 230) has given the best values for the other cruises. Ewing (1937) has shown that the negative trough continues around the Windward Islands to Trinidad. None of the Barracuda measurements are considered in our section as all these values are too far to the east. The U. S. Coast and Geodetic Survey has made four gravity stations on Puerto Rico. Heiskanen (1939, p. 92) also lists these stations. The only isostatic anomaly computed for all the various stations is the Hayford isostatic anomaly with a depth of compensation of 113.7 km. Table 1 shows the various gravity stations, the water depth, the free-air anomalies, the Hayford isostatic anomalies, and the available Airy local isostatic anomalies for a crustal thickness of 30 km. Utilizing either the Hayford or the Airy theory, it is obvious that the trench is not in isostatic equilibrium. In fact the Airy hypothesis (T = 30 km) quickly leads to impossible situations. If one uses the local Airy hypotheses with T = 30 km, water density 1.03, crustal density 2.67, and subcrustal density 3.27, one finds that the subcrustal material lies at the ocean floor for water depths of 8 km and above the ocean floor for greater water depths. For T = 20 km this dilemma arises at water depths of 5.7 km, just slightly more than the mean ocean depth. If a density differential of 0.3 is chosen between the crustal density and the subcrustal density, these figures become 4.7 km for T = 30 km and 3.2 km for T = 20 km. Regional isostatic assumptions will not modify this picture appreciably.

Since isostatic assumptions of density distribution do not fit the data, we can discard the isostatic anomalies and go back to the free-air anomalies, trying other density distributions to account for the observed anomalies.

Structure Deducted from Seismic and Gravity Data

We have utilized the seismic evidence to form most of our picture, choosing a density of 2.30 for the sediments. We have chosen this density as about the best average for water-filled

sediments (Birch *et al.*, 1942, p. 19–26). Although the sediments of the ocean bottom are subjected to great hydrostatic pressure, it is improbable that this causes compaction as a load of overburden does (Hedberg, 1936). While the thickness of overburden in the trench is sufficient to effect compaction, this sediment was probably collected so quickly that the compaction process has had insufficient time to operate, so that a density of 2.30 is probably a good figure.

We have made our computations for rectangular blocks extending to infinity in the directions parallel to the axis of the trench. This is quite as good a computation as the quantity of gravity data merits. We hope to obtain many gravity observations on a profile across the trench in the near future, and it would be more appropriate to use more sophisticated computations at that time.

The seismic data does not provide the necessary evidence for the thickness of the 5.5- to 6.0-km/sec layer near Puerto Rico, so we have adjusted this thickness until the gravity data matched fairly well. Likewise, we have neither measured the thickness of the sediments in the trough nor detected any underlying layers. We have assumed a crustal thickness like that found under the ocean north of the trench and have adjusted the sedimentary thickness until we obtained the best fit for the gravity data. Figure 2B shows the assumed structure and densities used in the computation of the best fit to the gravity data. Figure 2C shows the observed free-air anomalies plotted as points, and the results of our computation shown as a continuous curve. It is obvious that the fit is quite satisfactory.

It is interesting to note that a 6-km thickness of sediments is required to fit the gravity data for the trench and that the subsedimentary layer near Puerto Rico must be 12 km thick. If the sedimentary density were higher, the thickness of sediments would have to be greater. No unusual downbulge is required under the trough to explain the gravity anomalies.

Figure 2D shows without vertical exaggeration the structural section deduced from the seismic and gravity data. The existence of the basaltic (7.0–7.2 km/sec) layer under the trench and under Puerto Rico is purely hypothetical and is based on the notion that these were formerly normal oceanic crusts. The bulk of the material under Puerto Rico with a velocity of 5.5 to 6.0 km/sec is probably metamorphic rocks like those visible on the surface of Puerto Rico.

The negative gravity anomaly belt may be explained without demanding any significant crustal shortening or lengthening (Fig. 2D). Omitting the basaltic layer under the trench or doubling its thickness there would give a section that would not seriously contradict the seismic data. Owing to the slight thickness of the basaltic layer in undisturbed ocean-bottom crust, these limits of thickness would be consistent with crustal deformation ranging from 100 km of extension to an equally great shortening. In other words the solution offered in Figure 2 imposes no demands for large relative horizontal displacements of crustal segments. In fact it indicates that study of the gravity anomalies cannot be expected to give information about such displacements. The existence of strong belts of negative gravity anomalies near the southern and western borders of the Caribbean Sea (indicated by Vening Meinesz *et al.*, (1934), p. 156, 230 and Ewing (1937), but detailed in our recent unpublished measurements) make it very awkward to explain the large negative gravity anomalies along the northern and eastern boundaries by large horizontal crustal displacements.

Since seismic-refraction measurements permit only a very slight thickness of low-density crustal rocks either north or south of Puerto Rico, a thick prism of sediments seems to offer the only possibility of explaining the observed negative gravity anomalies.

We consider it of greatest importance to make additional seismic and gravity measurements in the West Indian area to obtain conclusions based entirely on positive and abundant data. It is obviously also important to obtain seismic-refraction measurements in the Pacific trenches —especially in the Marianas and the East Indies—to learn if the West Indian arc is typical.

Hypothesis on the Origin of the Island Arc and Some Consequences of It

The hypothesis is offered that the Puerto Rico trench with its rapidly accumulating sediments represents a first stage in the develop-

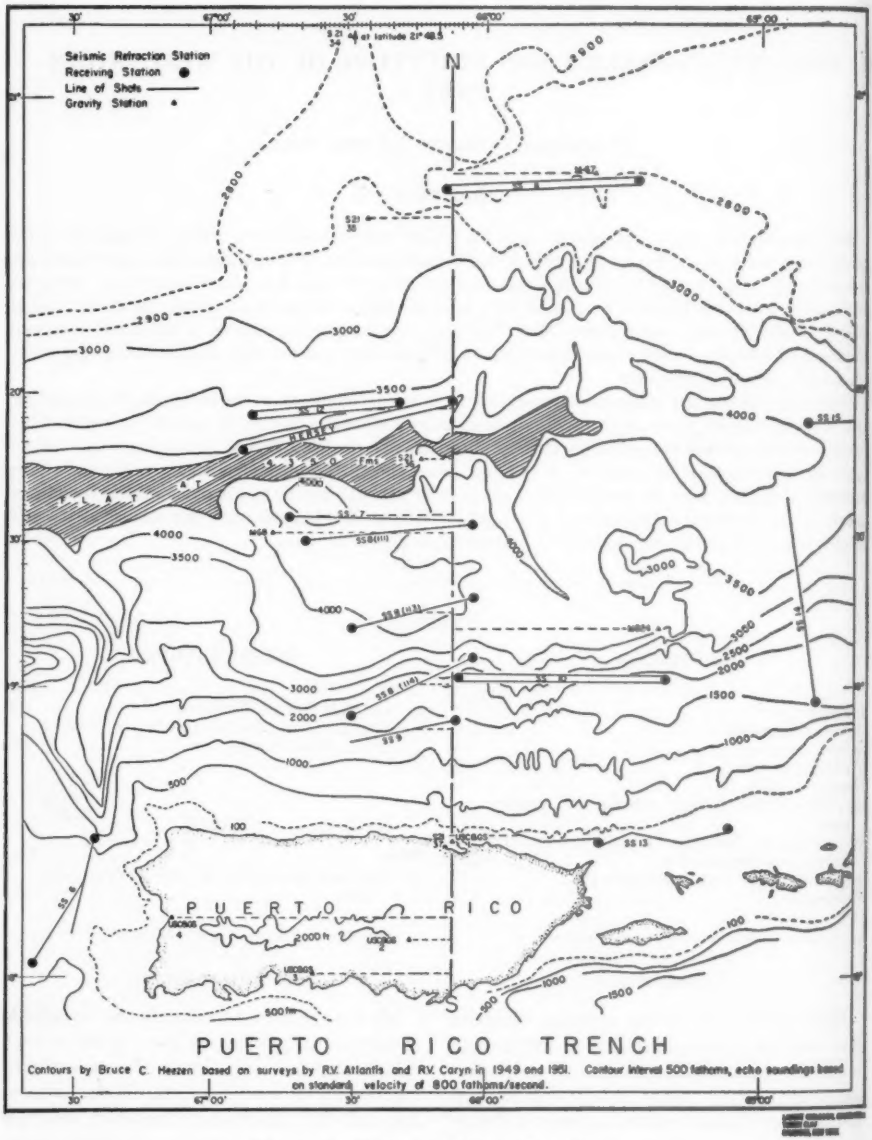

Seismic Refraction Station
Receiving Station
Line of Shots
Gravity Station
N
PUERTO RICO
PUERTO RICO TRENCH
Contours by Bruce C. Heezen based on surveys by R.V. Atlantis and R.V. Caryn in 1949 and 1951. Contour interval 500 fathoms, echo soundings based on standard velocity of 800 fathoms/second.

FIGURE 1.—CHART OF PUERTO RICO AREA

SEISMIC RESULTS

Six seismic-refraction measurements were made in the spring of 1951 in and near the trench north of Puerto Rico. These together with a measurement made by Hersey in 1949 are shown in Figure 1 superimposed on the best bathymetric chart available for publication, H.O. 5487. The locations of the gravity stations observed by Vening Meinesz and the U. S. Coast and Geodetic Survey are also shown. A

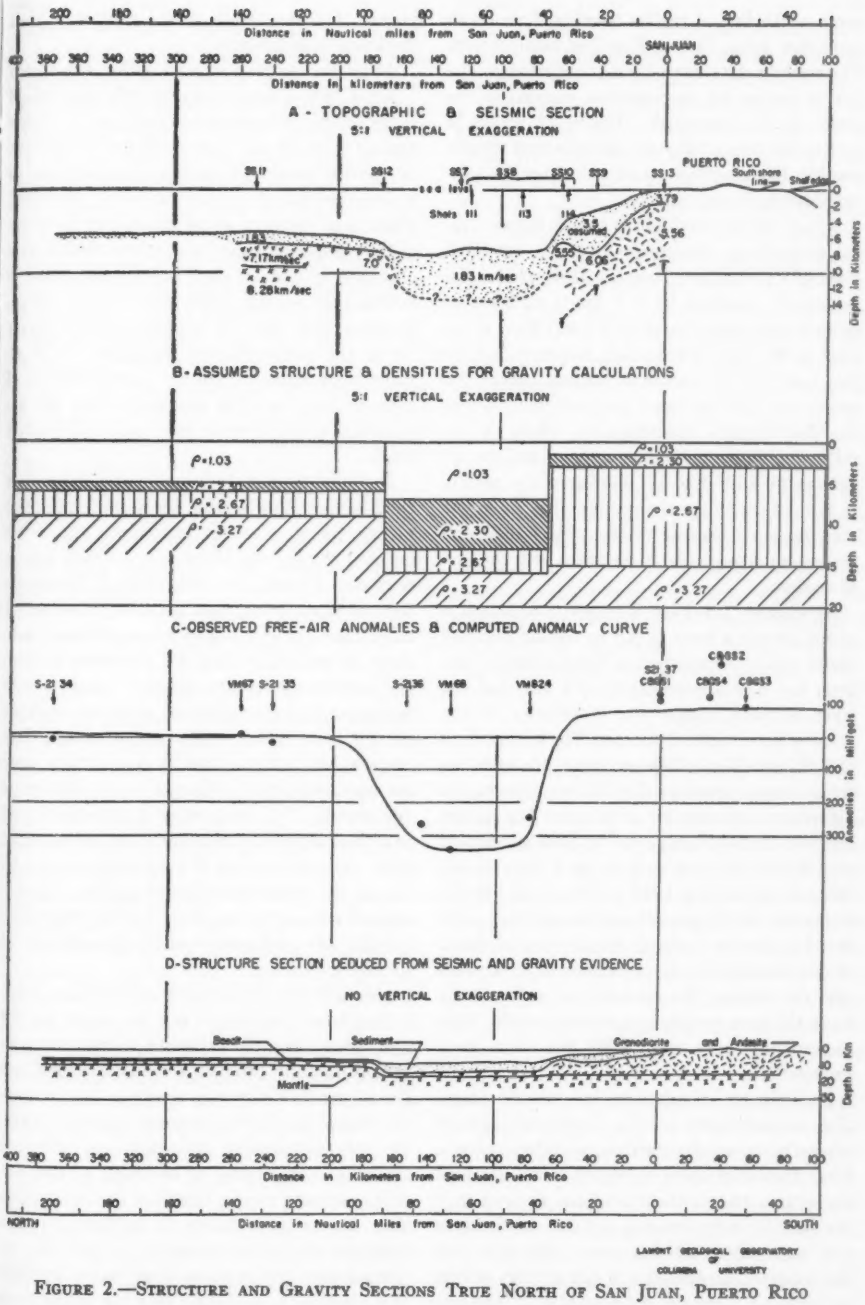


FIGURE 2.—STRUCTURE AND GRAVITY SECTIONS TRUE NORTH OF SAN JUAN, PUERTO RICO

north-south line of profile trending true north from San Juan, Puerto Rico has been chosen. The various data have been transferred to the line of section by moving them parallel to the strike of the topography. Figure 2A shows in section the topography and the observed seismic results. The observed sound velocities are indicated beneath each station.

S.S.-11 80 km north of the trench shows typical ocean-basin structure—*i.e.*, a 3 km-thick layer of 7.17 km/sec above the Mohorovičić discontinuity, overlain by a 1 km-thick layer of sediments in ocean depths of 5 km (Officer *et al.*, 1952, p. 801, 802; Ewing *et al.*, In press). S.S.-12 just north of the trench is similar, except the profile was not extended far enough to detect the Mohorovičić discontinuity. Hersey's results are, though fragmentary, essentially in agreement with S.S.-12. Similarly the results at S.S.-15 are in agreement, but they have not been included because of the large lateral displacement required to bring them into the line of section.

At stations S.S.-13, S.S.-9, and S.S.-10 south of the trench a layer of 5.5 to 6.0 km/sec was found under the sediments. The bottom of this layer has a ***minimum*** depth of 9 km, and the layer is quite comparable in velocity to the granitic layer under continents. The seismic data of S.S.-6 in Mona Passage have not been included here because of the obvious topographic difference from the line of section. The deeper layers in Mona Passage are in good agreement with S.S.-13. Seismic work in the Virgin Islands (Worzel and Ewing, 1948, p. 46) about 150 km to the east of the present section shows consolidated sediments (velocity 4.0 km/sec) of about .45 km thickness overlying a layer of 5.6 km/sec near the islands. The consolidated sedimentary layer thickens greatly toward the trench. This corresponds very well with the data near Puerto Rico.

Preliminary calculations on seismic-refraction measurements in the Caribbean indicate typical ocean-basin structure, as outlined above. Since the ocean-basin structure found north of the trench and in the Caribbean showed that the sialic layer was missing and the simatic layer was quite thin, making great difficulties for the accepted explanation of the gravity deficit in terms of a crustal down-buckle, it was expected that the trench would contain a great thickness of sediments.

At all the stations located within the trench —S.S.-7, S.S.-8 (shots 111 and 113), and S.S.-14 —the seismic measurements gave positive information only for the unconsolidated sediments; i.e. neither basement rocks nor consolidated or semi-consolidated sediments were detected. There is no question about the proper operation of the equipment since satisfactory results were obtained with it on days before, after, and in between the several observations in the trench. In more than 100 sea seismic stations, layers below the unconsolidated sediments have always been observed. On any interpretation of these results it must be concluded that the unconsolidated sediments are many kilometers thick.

Inability to map basement rocks beneath a thick sedimentary section by seismic-refraction measurements is not an uncommon experience on land or near the shore line. In such places, semi-consolidated or consolidated sediments were always found, and none were buried by more than about 1 km of unconsolidated sediment. In general at deep-sea refraction stations the entire sedimentary column (about 1 km) is apparently unconsolidated, although evidence on this point is inherently hard to obtain. It is remarkable that no velocities greater than those for unconsolidated sediments were observed in the trench. The compaction commonly found in a thick sedimentary column has not occurred here, perhaps because of very rapid accumulation of the sediment or because of the dependence of compaction on factors other than mere weight of overburden—*e.g.*, percolation of ground water.

One estimate of the minimum thickness of unconsolidated sediments can be made on the assumption that the explosive charge was sufficiently large to produce a readable signal, but the longer-shot distances were too short to permit reception of the basement-refracted wave. On this assumption the minimum sediment thickness is computed to be about 12 km. Another estimate can be based on the assumption that the great thickness of sediments causes sufficient absorption of energy to make the refracted wave too weak for observation. In effect this amounts to declaring that the longer shots

of each profile must be excluded from the calculation of minimum depth. For many reasons it is impossible to make precise calculations of the absorption. A rough estimate may be made as follows: For comparable conditions at a typical deep-sea station the longest horizontal travel for the refracted wave travelling in unconsolidated sediments is 8–16 km. Assuming vertical and horizontal absorption are approximately equal, and that no other factors enter, the minimum thickness of sediments in the trench would be 4–8 km. Additional seismic-refraction measurements in the trench are needed.

Barbados, which lies on a continuation of the gravity minimum associated with the Puerto Rico trench, was investigated in 1944 (Worzel and Ewing, 1948, p. 48), and the basement rocks were not detected. It was concluded that about 1 km of unconsolidated sediments and 1 km of semiconsolidated sediments covered more than 3 km of consolidated sediments. This is consistent with the idea that the sediments were laid down under conditions now prevailing in the Puerto Rico trench.

Turbidity Currents

Ericson *et al.* (1951; 1952) and Heezen and Ewing (1952) have presented abundant evidence of the widespread operation, in the late Cenozoic oceans, of turbidity currents capable of carrying sediments hundreds of miles. The types of evidence of turbidity currents presented are sands containing shallow-water species of Foraminifera interbedded with oozes of abyssal facies, graded bedding in the turbidity-deposited layers with sharp lower boundaries, high-carbonate gray clays repeatedly alternating with red clays in basins, particles of shallow-water flora and fauna of Pleistocene and Recent age found in the turbidity-deposited layers at depths greater than 2000 fathoms, mineral and faunal similarity of deep-sea sands and continental-shelf sediments, and finally the flat floors of the deepest depressions being formed of numerous sedimentary layers exhibiting many alternations of colors and sediments.

The two cores (±10 feet long) taken in the flat plain on the bottom of the Puerto Rico trench clearly show deposition of much of the sediment by turbidity currents (Ericson *et al.*, 1952, p. 496–498, 509). These cores consist largely of thick graded calcareous sands containing shallow-water fauna and flora indicating a very large contribution of sediment to this trough from the surrounding shallow-water areas. Red clays are interbedded with the calcareous sands. Three cores on the north wall of the trench are normal deep-sea red clay throughout. Two cores on the south wall are essentially all clay, showing evidence of slumping. One is Pleistocene, the other Miocene. All of these cores are from depths so great that they would be expected to be red clay throughout if the deposition occurred in the absence of transportation along the bottom. Available bathymetric data indicate that a continuous ridge caps the north and east walls of the trench, cutting it off from turbidity currents from the main ocean basin. It is improbable that the barrier has been effective throughout the existence of the trench. Hence, if one grants the possibility of turbidity currents from the main ocean basin to the trench, it is reasonable to suppose that much of the sedimentary column contains a much smaller percentage of calcium carbonate than that reached by the coring tubes.

Gravity Results

Reconsideration of Vening Meinesz' (1934, p. 117–133; 1948, p. 25–90) and Hess's (1933; 1938, p. 71–86; 1951) hypothesis of the sialic crust down-buckling to form a local thickening of the sialic crust (tectogene) is required because of the seismic-refraction evidence that the sialic crust is absent under the Atlantic Ocean and under the Caribbean Sea. Vening Meinesz required a cross section of 1200 km² for a tectogene with a density contrast of 0.6 at the base of a 25- or 30-km crust, to produce the observed anomaly, corresponding to a crustal shortening of 50 or 40 km respectively. If the 3- to 5-km-thick layer found by seismic-refraction methods to overlie the Mohorovičić discontinuity and to underlie the sediments of the ocean floor was the source of the thickened section, the shortening required would be 250 to 500 km. The 1200 km² block calculated by Vening Meinesz was for the Java-Christmas Island profile where the largest isostatic anomaly was −130 mgals,

whereas the largest isostatic anomaly reported by Vening Meinesz (1948, p. 230) for the trench north of Puerto Rico was −226 mgals (the largest negative anomaly reported to date). Thus more than 500 km of shortening would be required to produce the anomaly. Hess (1933, p. 31) prefers a crust with a density 0.3 less than the material beneath. This figure doubles the required crustal shortening.

Table 1.—Isostatic Anomalies in the Puerto Rico Area

Station		Water depth (Meters)	Free-air anomaly (Mgals)	Hayford isostatic anomaly (Mgals)	Airy local isostatic anomaly (T = 30 km)
S-21	34	5569	−18	−1	
	35	5340	+21	+31	
	36	7699	−264	−125	
	37	9	+129	−19	
Vening Meinesz	67	5510	+9	+7	+7
	68	8040	−341	−193	−292
	824	7370	−244	−226	−222
Elevations					
U. S. Coast & Geodetic Survey	1	+8	+112	−43	On land
	2	+648	+220	+23	
	3	+14	+91	−48	
	4	+16	+122	−20	

The location of the gravity stations in the Puerto Rico area is shown in Figure 1. Vening Meinesz first discovered the negative gravity trough associated with this topographic trench on his world-circling cruise of Hr. Ms. K-XIII. On the cruise of the S-21 he made additional observations and again on Hr. Ms. 0 13. Heiskanen (1939, p. 98) has given the best values for the S-21 data, while Vening Meinesz (1948, p. 156, 230) has given the best values for the other cruises. Ewing (1937) has shown that the negative trough continues around the Windward Islands to Trinidad. None of the Barracuda measurements are considered in our section as all these values are too far to the east. The U. S. Coast and Geodetic Survey has made four gravity stations on Puerto Rico. Heiskanen (1939, p. 92) also lists these stations. The only isostatic anomaly computed for all the various stations is the Hayford isostatic anomaly with a depth of compensation of 113.7 km. Table 1 shows the various gravity stations, the water depth, the free-air anomalies, the Hayford isostatic anomalies, and the available Airy local isostatic anomalies for a crustal thickness of 30 km. Utilizing either the Hayford or the Airy theory, it is obvious that the trench is not in isostatic equilibrium. In fact the Airy hypothesis (T = 30 km) quickly leads to impossible situations. If one uses the local Airy hypotheses with T = 30 km, water density 1.03, crustal density 2.67, and subcrustal density 3.27, one finds that the subcrustal material lies at the ocean floor for water depths of 8 km and above the ocean floor for greater water depths. For T = 20 km this dilemma arises at water depths of 5.7 km, just slightly more than the mean ocean depth. If a density differential of 0.3 is chosen between the crustal density and the subcrustal density, these figures become 4.7 km for T = 30 km and 3.2 km for T = 20 km. Regional isostatic assumptions will not modify this picture appreciably.

Since isostatic assumptions of density distribution do not fit the data, we can discard the isostatic anomalies and go back to the free-air anomalies, trying other density distributions to account for the observed anomalies.

Structure Deducted from Seismic and Gravity Data

We have utilized the seismic evidence to form most of our picture, choosing a density of 2.30 for the sediments. We have chosen this density as about the best average for water-filled

sediments (Birch *et al.*, 1942, p. 19–26). Although the sediments of the ocean bottom are subjected to great hydrostatic pressure, it is improbable that this causes compaction as a load of overburden does (Hedberg, 1936). While the thickness of overburden in the trench is sufficient to effect compaction, this sediment was probably collected so quickly that the compaction process has had insufficient time to operate, so that a density of 2.30 is probably a good figure.

We have made our computations for rectangular blocks extending to infinity in the directions parallel to the axis of the trench. This is quite as good a computation as the quantity of gravity data merits. We hope to obtain many gravity observations on a profile across the trench in the near future, and it would be more appropriate to use more sophisticated computations at that time.

The seismic data does not provide the necessary evidence for the thickness of the 5.5- to 6.0-km/sec layer near Puerto Rico, so we have adjusted this thickness until the gravity data matched fairly well. Likewise, we have neither measured the thickness of the sediments in the trough nor detected any underlying layers. We have assumed a crustal thickness like that found under the ocean north of the trench and have adjusted the sedimentary thickness until we obtained the best fit for the gravity data. Figure 2B shows the assumed structure and densities used in the computation of the best fit to the gravity data. Figure 2C shows the observed free-air anomalies plotted as points, and the results of our computation shown as a continuous curve. It is obvious that the fit is quite satisfactory.

It is interesting to note that a 6-km thickness of sediments is required to fit the gravity data for the trench and that the subsedimentary layer near Puerto Rico must be 12 km thick. If the sedimentary density were higher, the thickness of sediments would have to be greater. No unusual downbulge is required under the trough to explain the gravity anomalies.

Figure 2D shows without vertical exaggeration the structural section deduced from the seismic and gravity data. The existence of the basaltic (7.0–7.2 km/sec) layer under the trench and under Puerto Rico is purely hypothetical and is based on the notion that these were formerly normal oceanic crusts. The bulk of the material under Puerto Rico with a velocity of 5.5 to 6.0 km/sec is probably metamorphic rocks like those visible on the surface of Puerto Rico.

The negative gravity anomaly belt may be explained without demanding any significant crustal shortening or lengthening (Fig. 2D). Omitting the basaltic layer under the trench or doubling its thickness there would give a section that would not seriously contradict the seismic data. Owing to the slight thickness of the basaltic layer in undisturbed ocean-bottom crust, these limits of thickness would be consistent with crustal deformation ranging from 100 km of extension to an equally great shortening. In other words the solution offered in Figure 2 imposes no demands for large relative horizontal displacements of crustal segments. In fact it indicates that study of the gravity anomalies cannot be expected to give information about such displacements. The existence of strong belts of negative gravity anomalies near the southern and western borders of the Caribbean Sea (indicated by Vening Meinesz *et al.*, (1934), p. 156, 230 and Ewing (1937), but detailed in our recent unpublished measurements) make it very awkward to explain the large negative gravity anomalies along the northern and eastern boundaries by large horizontal crustal displacements.

Since seismic-refraction measurements permit only a very slight thickness of low-density crustal rocks either north or south of Puerto Rico, a thick prism of sediments seems to offer the only possibility of explaining the observed negative gravity anomalies.

We consider it of greatest importance to make additional seismic and gravity measurements in the West Indian area to obtain conclusions based entirely on positive and abundant data. It is obviously also important to obtain seismic-refraction measurements in the Pacific trenches —especially in the Marianas and the East Indies—to learn if the West Indian arc is typical.

Hypothesis on the Origin of the Island Arc and Some Consequences of It

The hypothesis is offered that the Puerto Rico trench with its rapidly accumulating sediments represents a first stage in the develop-

ment of a landmass like the Greater Antilles. Without specifying the forces that cause the deep depressions to form or the sources of heat that elevate the temperature of magmas and lavas in general, we point out that granitization of a body of sediments like that now collecting in the Puerto Rico trench assuming that the bulk of the sediment is comparable to deep-sea sediments in general—*e.g.*, red clay, or like the oceanic formation on Barbados) would produce a mass of sialic rock that could be expected to rise and form an island arc when the force which caused the trench was relaxed. The Barbados rise and the Lesser Antillean arc would then represent intermediate stages.

Among the additional consequences of this hypothesis may be mentioned the following: (1) Andesitic lavas result from the contamination of basaltic-magmas, by the sediments during the granitization of the sediments. (2) Most of the water released by andesitic volcanoes is not juvenile water, but merely the sea water entrapped when the sediments were deposited.

Implications for Continental Growth

The formation of trenches at or near the edges of continents, the rapid filling of the trenches by sediments, the alteration of the sediments to sialic rocks, and the rise of these rocks to form island chains provide a means for confining continental debris to the proximity of continents despite the action of turbidity currents. Such trenches at present almost completely ring the Pacific Ocean, probably reducing the role of turbidity-current deposition in mid-ocean sedimentation far below that which it plays in the Atlantic. Types of present-day trenches range from extreme oceanic types such as the Marianas to extreme continental types such as the west coast of Central and South America. All these types have probably existed throughout much of geologic time and have probably bordered nearly every coast.

The total depth of deep-sea sediment is somewhat less than 1 km. Only by impounding most of the continental debris near continental margins and by transfer of sediments from the ocean basins to the continental margins can such a small thickness be explained. The alternatives are either an average rate of sedimentation less than 1 mm per 500 years or a process of metamorphism of sediments at a depth of less than 1 km beneath the ocean floor.

A detailed consideration of the role of deep-sea sediment in continental growth will be deferred, but it is clear that the process we have described for the West Indian arc would greatly facilitate the growth of continents from an original crust of basaltic rocks as suggested by A. C. Lawson (1932), Marshall Kay (1951), J. Tuzo Wilson (1949), and others.

References Cited

Birch, F., Schairer, J. F., and Spicer, H. C. (Editors), 1942, Handbook of physical constants: Geol. Soc. America Spec. Paper 36, 325 p.

Ericson, D. B., Ewing, Maurice, and Heezen, B. C., 1951, Deep-sea sands and submarine canyons: Geol. Soc. America Bull., v. 62, p. 961–965.

—— 1952, Turbidity currents and sediments in North Atlantic: Am. Assoc. Petroleum Geologists Bull., v. 36, p. 489–511.

Ewing, Maurice, 1937, Gravity measurements on the U.S.S. Barracuda: Am. Geophys. Union Trans., 18th Ann. Mtg., p. 66–69.

—— 1952, Seismic investigations in great ocean depths: Assoc. Oceanogr. Phys., Proces-Verbaux no. 5, p. 135–136.

Ewing, Maurice, Sutton, G. H., and Officer, C. B., Jr., In press, Seismic refraction measurements in the Atlantic Ocean. Part VI, Typical deep stations, North American Basin: Seismol. Soc. America Bull.

Hedberg, Hollis B., 1936, Gravitational compaction of clays and shales: Am. Jour. Sci., v. 31, p. 241-287.

Heezen, B. C., and Ewing, Maurice, 1952, Turbidity currents and submarine slumps, and the 1929 Grand Banks earthquake: Am. Jour. Sci., v. 250, p. 849–873.

Heiskanen, W., 1939, Catalogue of the isostatically reduced gravity stations: Isostatic Inst. of Internat. Assoc. of Geodesy Pubs., no. 5, 139 p.

Hess, H. H., 1933, Interpretation of geological and geophysical observations in Navy Princeton Gravity Expedition to the West Indies in 1932: U. S. Hydrog. Office, p. 26–38.

—— 1938, Gravity anomalies and island arc structure with particular reference to the West Indies: Am. Philos. Soc. Proc., v. 79, no. 1, p. 71–96.

—— 1951, Comment on mountain building: Am. Geophys. Union Trans., v. 32, no. 4, p. 528–531.

Kay, Marshall, 1951, North American geosynclines: Geol. Soc. America Mem. 48, 143 p.

Lawson, A. C., 1932, Insular arcs, foredeeps, and geosynclinal seas of the Asiatic coast: Geol. Soc. America Bull., v. 43, p. 353–381.

Officer, C. B., Ewing, Maurice, and Wuenschel,

P. C., 1952, Seismic refraction measurements in the Atlantic Ocean. Part IV: Bermuda, Bermuda Rise, and Nares Basin: Geol. Soc. America Bull., v. 63, p. 777–808.

Vening Meinesz, F. A., 1948, Gravity expeditions at sea 1923–1938: Delftsche Uitgevers Maatschappil-Delft, v. IV, 233 p.

——1954, Indonesian Archipelago, a geophysical study: Geol. Soc. America Bull., v. 65, p. 143–164.

Vening Meinesz, F. A., Umbgrove, J. H. F., and Kuenen, Ph.H., 1934, Gravity expeditions at sea 1923–1932: N. V. Tech. Boekhandel en Brukkerij J. Waltman Jr. Delft, v. II, 208 p.

Wilson, J. Tuzo, 1949, Origin of continents and Precambrian history: Royal Soc. Canada Trans., 3d ser., vol. 43, sec. 4, p. 157–184.

Worzel, J. L., and Ewing, Maurice, 1948, Explosion sounds in whallow water, *in* Worzel, J. L., Ewing, Maurice, and Pekeris, C. L., Propagation of sound in the ocean: Geol. Soc. America Mem. 27, p. 1–53.

LAMONT GEOLOGICAL OBSERVATORY (COLUMBIA UNIVERSITY), PALISADES, N. Y.

MANUSCRIPT RECEIVED BY THE SECRETARY OF THE SOCIETY, JANUARY 12, 1953

LAMONT GEOLOGICAL OBSERVATORY CONTRIBUTION NO. 99

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 65, PP. 175-190, 7 FIGS., 2 PLS. FEBRUARY 1954

GEOLOGY OF ALLUVIAL FANS IN SEMIARID REGIONS

BY ERICH BLISSENBACH

ABSTRACT

An alluvial fan is a body of detrital sediments built up at a mountain base by a mountain stream. Bold relief is essential, moderately arid to semiarid climate favorable for the development of fans. The depositing agents are sheet floods, stream floods, and streams. Compound alluvial fans result from lateral coalescence of single fans.

Development of alluvial fans is affected by changes in the course of a cycle, varying base level, climatic changes, tectonic movements, and slumping of fan deposits. Telescoped or superimposed structure may be developed.

Fan deposits are arkosic or graywacke. Sorting and roundness of particles range widely. The matrix is primary or secondary. In general alluvial-fan deposits are stratified. Channel cut-and-fill is pronounced. Individual strata in fans are up to 20 feet thick. Particles in stream deposits are imbricated.

Talus-slope deposits at the apex of a fan and floodplain deposits at its base can be separated from those of an alluvial fan by particle sizes, angularity and orientation of fragments, sorting, and original dip of strata. Mudflow deposits in an alluvial fan indicate certain climatic conditions during its formation.

Many ancient fan deposits may have escaped recognition because of the common misconception that fan deposits are necessarily unstratified, composed of angular fragments, poorly sorted, and without distinctive sedimentary structures.

CONTENTS

TEXT

	Page
Introduction	176
Purpose of investigation	176
Areas of field study	176
Acknowledgments	176
Geology of alluvial fans	176
Definitions	176
General statement	176
Process of forming alluvial fans	177
Late-stage development	179
Changes in normal course of a cycle	179
Effects of varying base level	179
Effects of climatic changes	180
Effects of tectonic movements	180
Effects of slumping of unconsolidated fan material	181
Secondary alluvial fans	181
Geology of alluvial-fan deposits	181
Facies	181
Particle sizes and particle-size distribution	182
Composition	182
Sorting	183
Roundness	183
Sphericity	184
Matrix	184
Cement	185
Color	185
Porosity	185
Permeability	185
Sedimentary structures	185
Organic contents	187
Ancient alluvial-fan deposits	187
Occurrence	187
Relative abundance	187
Criteria for recognition of ancient alluvial-fan deposits	188
References cited	189

ILLUSTRATIONS

Figure	Page
1.—Compound alluvial fan formed by lateral coalescence of single alluvial fans	177
2.—Alluvial fan with telescope structure	180
3.—Superimposed alluvial fan	181
4.—Pseudo-telescope structure (modified after Gilbert)	181
5.—Distribution of maximum particle sizes and surface angles along a radial profile on an alluvial fan of the Santa Catalina Mountains, Arizona	182
6.—Distribution of roundness and sphericity of alluvial fan particles along the same radial profile as Figure 5. Base of Santa Catalina Mountains, Arizona	184
7.—Characteristic sedimentary structures in an alluvial fan of the Santa Catalina Mountains, Arizona. Main direction of transport from right to left	186

Plate	Facing page
1.—Alluvial fans	186
2.—Alluvial-fan deposits	187

Introduction

Purpose of Investigation

This paper gives the results of a comprehensive study of alluvial fans. Though detailed field work was carried out only in the southwestern United States, most of the data may be applied to alluvial fans in other areas where similar conditions prevail. Thus, through detailed study of recent alluvial-fan deposits, geologists may be aided in determining whether ancient rocks were formed under similar conditions.

Areas of Field Study

Alluvial fans studied in southern Arizona include those at the southern base of the Santa Catalina Mountains, north and northeast of Tucson, Arizona, and fans of the Tucson Mountains, west and northwest of Tucson, Arizona.

Investigations in the Mammoth area, Arizona, were carried out on alluvial fans on both sides of an intermontane valley of the Black Hills, Arizona, west of Mammoth.

In northern Arizona alluvial fans were studied along the Aubrey Cliffs, northwest of Seligman, Arizona, and along the Bright Angel Trail, north of the Grand Canyon Village, Arizona.

Studies were also carried out on alluvial fans in southern California and southern New Mexico.

Certain alluvial fans were studied in every detail, whereas only certain properties were examined in other fans.

Acknowledgments

The aid and counsel of Professor Edwin D. McKee of the University of Arizona, under whose guidance this work was undertaken, is sincerely appreciated.

Geology of Alluvial Fans

Definitions

An ***alluvial fan*** is a body of detrital sediments built up by a mountain stream at the base of a mountain front. It develops because all streams tend to form a graded course. A talus slope differs from an alluvial fan in that it is built up predominantly by gravitational sliding.

The *apex* of an alluvial fan develops at the point where the stream emerges from the mountain. It is the point of highest elevation on the alluvial fan. *Fanhead* applies to the area on the alluvial fan close to the apex; *midfan* designates the area between the fanhead and the outer, lower margins of the fan. *Base* of an alluvial fan is the term applied to the outermost or lowest zone of the fan.

If the fanhead area reaches far into a mountain canyon the term *fan-bay* (Davis, 1938, p. 1374) is applied to it. *Fan-mesa* (Eckis, 1928, p. 243–244) is the term applied to an alluvial fan remnant left standing in the process of degradation of a fan.

General Statement

The occurrence of alluvial fans has been reported from all continents. However, due to conspicuous development and easy accessibility, alluvial fans in California have been studied in more detail than any others (Lawson, 1913, p. 332; Eckis, 1928, p. 232–246; Krumbein, 1937, p. 586–594; Buwalda, 1951, p. 1491).

Investigators generally agree that an alluvial fan resembles geometrically the segment of a cone. From the apex of the fan the surface dips toward the base in which direction the angles of dip gradually become flatter. Thus, a radial profile through the fan is concave upward; a profile at right angles convex. The steepest angle of dip on the alluvial fan is encountered close to its apex.

The angle of dip of the fan surface rarely exceeds 10° (Dana, 1894, p. 194–195; Eckis, 1928, p. 223; Scott, 1932, p. 269; Eardley, 1938, p. 1408). Some authors report no angle greater than 5° or 6° (Lawson, 1915, p. 25; Vaughan, 1922, p. 341). The maximum angles of slope of alluvial fans studied by the writer range from 5° observed on fans of the Aubrey Cliffs, Arizona, to 9° on fans of the Black Hills of Arizona.

Surface angles greater than 5° are characteristic of the upper half or more of small alluvial fans with a radial extent of a few hundred feet like those of the Black Hills of Arizona.

Large alluvial fans such as those of the Santa Catalina Mountains, Arizona, with a radial extent of about 4 miles, exhibit surface angles greater than 5° only within the upper one twentieth or less of their extent.

In geologic literature small alluvial fans are commonly called alluvial cones. As no definition specifies the difference between an alluvial fan and a cone, this classification is not followed. Instead the terms steep, gentle, and flat angles of dip are proposed to denote the degree of slope of alluvial fans or parts of them.

Because angles of dip greater than 5° are rare these are called steep. Angles of dip between 2° and 5° are termed gentle, and sloping angles below 2° flat. On the basis of these definitions the slope of the surface of a particular alluvial fan can be described as follows:

Alluvial fans of the Santa Catalina Mountains with a radial extent of about 4 miles have steep slopes in the uppermost 0.2 mile of the radial extent of the fans. In the adjoining 0.7 mile surface angles are gentle, while the remaining part, about 3.1 miles, has flat angles of dip.

The radius of an alluvial fan may be as great as 40 miles under exceptional conditions (Grabau, 1913, p. 584). The radii of alluvial fans studied by the writer range from 4 miles at the Catalina Mountains, Arizona, to about 500 feet in the Black Hills, Arizona.

By lateral coalescence of single alluvial fans, a *compound alluvial fan* may result (Miller, 1926, p. 164–166) (Fig. 1; Pl. 1, fig. 1). The compound alluvial fan is equivalent to the alluvial piedmont slope (Lahee, 1941, p. 101) and the "bajada" (Blackwelder, 1931, p. 136).

Alluvial fans may be 1000 or more feet in thickness (Eckis, 1928, p. 224). In general, the thickness of alluvial-fan deposits is greatest at the apex of a fan (Scott, 1932, p. 269). Under exceptional conditions, however, a fan may be thicker at the base than at its apex. This is illustrated in an alluvial fan of the Santa Catalina Mountains, Arizona; the thickness of the fan deposits increases from between 0 and 100 feet at the apex to 200 or 300 feet at the base of this fan.

Alluvial fans occur in areas of bold relief, and their development is most conspicuous under moderately arid to semiarid conditions. The influence of bold relief seems obvious as only under such conditions will there be profound erosion and transportation together with a strong tendency for deposition as the mountain streams reach areas of low gradient.

Practically all reductions of rock slope and contributions to alluvial embankment are

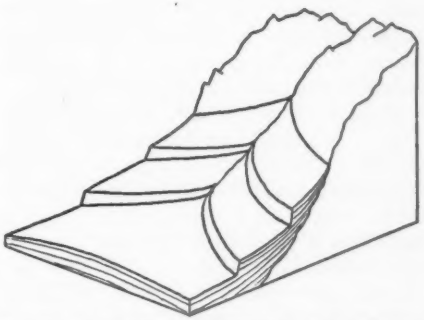


FIGURE 1.—COMPOUND ALLUVIAL FAN FORMED BY LATERAL COALESCENCE OF SINGLE ALLUVIAL FANS

affected by brief and infrequent periods of downpour (Lawson, 1915, p. 28). This type of precipitation is characteristic of areas with arid and semiarid climate. Under humid conditions also alluvial fans are developed as seen along the Alps and the Himalayas. Fans formed in humid environment commonly are flatter than those of arid environment owing to the abundance of running water which favors the development of gentler gradients. Under conditions of extreme aridity, mountains are buried under their own debris (Lawson, 1906, p. 449), and the deposits formed are in the nature of talus slopes.

Thus, the climatic conditions most favorable for the development of alluvial fans appear to range from moderately arid to semiarid. The annual precipitation in the region of an alluvial fan studied by Eckis (1928, p. 225) averages about 17 inches. Alluvial fans studied by the writer in Arizona receive a mean annual precipitation ranging from 10 to 19 inches.

Process of Forming Alluvial Fans

In the stage of deposition the surface of an alluvial fan shows a system of radiating channels focused in the main stream at the apex of the fan. This physiographic pattern is

referred to as a braiding stream system. Stream channels are distributaries. They commonly are shallow, entrenched but little beneath the gently sloping surface of an alluvial fan (Davis, 1898, p. 291).

Alluvial fans in process of degradation normally are cut by streams that tend to form deep and narrow channels. These streams are tributaries, and the resulting physiographic pattern constitutes a reversal of the braiding stream system focused in the apex of the fan.

Streams emerging from steep mountain canyons commonly are loaded with detrital material, especially if the times of flow are infrequent and separated by long, dry intervals. At the alluvial fan a pronounced tendency for deposition is immediately developed, because of a less steep gradient on the alluvial fan and the pervious nature of the alluvial-fan deposits which effects a continuous decrease in the volume of water.

Thus, fan channels frequently are silted up causing overflows and forming new distributaries. When one sector of a fan has been built up, the streams shift to another, lower section of the fan and build that up (Longwell, Knopf, and Flint, 1948, p. 88). The process is repeated again and again until the mountain stream and the alluvial fan have reached graded conditions.

In arid and semiarid regions short, violent flows normally constitute the transporting and depositing agents responsible for the development of alluvial fans. The depositing agents on alluvial fans may be classified into three types:

(1) Sheetfloods. These occur when an exceedingly large amount of water and detritus emerges from a mountain canyon. The flow, acting like a viscous medium, tends to spread out in the form of a sheet covering the alluvial fan or parts of it. Deep channels on alluvial fans tend to prevent sheetfloods, in the opinion of Davis (1938, p. 1348). Furthermore, he believed that the graded state of compound alluvial fans is effected largely by sheetfloods. A striking peculiarity of sheetfloods is the shortness in distance as well as in time of their flows. Great floods of this type may come only at intervals of decades or centuries (McGee, 1897, p. 108).

(2) Streamfloods. These are confined to definite channels on alluvial fans. They are formed where a large amount of water and detrital material emerges from a mountain canyon. Streamfloods may also form because channels on alluvial fans are too deep to allow a sheetflood to develop. The spasmodic and impetuous character of these floods is such that the term streamflood rather than streams is applied (Davis, 1938, p. 1347). The deposits of violent streamfloods tend to be identical with those of sheetfloods except that, instead of being blanket-shaped, they are linear in plan view.

(3) Streams. These are formed if both the amount of water and the quantity of detritus are less than the requirements for sheetfloods or streamfloods. A steady, rather than abundant supply and recharge of water from the mountains must be maintained. The action of streams is of minor significance in arid or semiarid climates for these climatic conditions do not favor the development of streams. In relatively humid regions, however, as in the Alps or the Himalayas, alluvial-fan deposits laid down by streams are of considerable magnitude.

Every possible gradation between these types of depositing agents must be expected on alluvial fans. Streamflood deposits exhibit gradations ranging from those formed during violent flows to others formed by floods of moderate intensity.

The deposits of sheetfloods and violent streamfloods tend to have a high percentage of mud-sized particles. The term mudflow deposits is commonly applied to these sediments, but this term should not be employed if the deposits can be ascribed to sheetflood or streamflood.

Any one alluvial fan may not show the effect of all three types of depositing agents. Furthermore, certain types of depositing agents may predominate at a particular stage in the development of an alluvial fan. Thus, the action of sheetfloods seems to be most pronounced when a graded stage has been reached—*i.e.*, with completion of the maturity of a fan (Davis, 1938, p. 1340).

The accumulation of flood waters in a mountain area may first give rise to a sheetflood and later, with decreasing recharge of water, change to streamfloods and finally to streams. The duration of sheetfloods is measured in terms of seconds or minutes (McGee, 1897, p. 101), that of streamfloods in minutes or

hours, and the flow of streams in hours or even days.

Pack (1923, p. 349–356) described a flood along the west front of the Wasatch Mountains, Utah, in which he recognized three stages of flooding: (1) torrential streamflood, (2) mudflow, (3) dwindling streamflood. These stages compare well with the classification of depositing agents adopted by the writer.

Alluvial fans studied by the writer in southwestern United States suggest that all three depositing agents—(1) sheetfloods, (2) streamfloods, and (3) streams—participated to a varying degree in their formation. In the Black Hills of Arizona, with a mean annual precipitation of 19 inches, mudflow deposits—*i.e.*, those formed by sheetfloods and violent streamfloods—constitute about 5 to 10 per cent of the fan deposits; moderate streamfloods or streams are responsible for the other 90 to 95 per cent. In the Tucson area with a mean annual precipitation of about 11 inches, mudflow deposits form from about 20 to 40 per cent of the alluvial fans, while streams and moderate streamfloods are responsible for the deposition of the other sediments in these fans.

Late-Stage Development

Changes in Normal Course of a Cycle

Until graded conditions are reached, erosion predominates along a mountain canyon, deposition on an alluvial fan. Upon development of a graded course in a mountain stream and an alluvial fan, erosion will reach over the apex of the fan and affect the fanhead area. Thus, the fanhead area undergoes dissection while a depositional tendency prevails farther downfan. The importance of such changes in the cycle of fan aggradation and degradation was first realized by Eckis (1928, p. 237).

Recession of a mountain front results in formation of a rock pediment normally covered by the fanhead part of an alluvial fan. Dissection of the fanhead area due to changes in the normal course of a cycle or to other causes may strip the rock pediment of its alluvial cover and expose it. A rock pediment may be an important indicator of the extent of an alluvial fan at its maturity.

The results attained by a single alluvial fan may also be developed on a compound type of fan. The normal change in a cycle, however, need not take place on adjacent units within a compound fan at the same time. Accomplishment of the maturity of a fan and the beginning of late-stage development is mainly a function of the size of the mountain stream and its gradient, factors which may vary considerably among different units of the same compound fan.

The compound fan at the southern base of the Santa Catalina Mountains, Arizona, is more heavily dissected along its eastern extent than in the west. In the absence of evidence for tectonic activities, the writer believes that the heavier erosion in the eastern part of the compound fan results from larger streams in the eastern part of the range whereby the maturity of fans and the beginning of late-stage development is accelerated. Slow development of alluvial fans and similarly slow changes in the cycle must be expected along the western extent of the compound fan where mountain streams are comparatively small.

Effects of Varying Base Level

Changes in the base level of a fan are commonly effected by a stream flowing along the fan base. Degradation of a fan base may also be caused by wind deflation as seen on some alluvial fans of the Aubrey Cliffs, Arizona (Blissenbach, 1952, p. 123).

Aggrading base level may be reflected in accelerated deposition on an alluvial fan. Under such conditions, the apex of a fan may migrate into the mountain canyon, and fan-bays may be developed.

Degrading base level results in erosion on an alluvial fan. Available evidence indicates that several of the fans studied by the writer have been affected by strong dissection due in part to the cutting in of a stream at the base of these fans.

The effects of the lateral swing of a river at the base of a fan is a combination of the effects of aggrading and degrading base level. The Colorado River has swung westward against the base of an alluvial fan, about 20 miles north of Needles. As a consequence of this undercutting, this alluvial fan is strongly dissected (Davis, 1938, p. 1349). In contrast, a

lateral swing of a river away from the fan base favors depositional tendencies on a fan, and the graded conditions may be restored.

A small alluvial fan of recent age, at the base of the compound fan of the Santa Catalina Mountains, Arizona, is built up of alternating layers of stream or streamflood and sheetflood deposits. Development of this fan is controlled in part by the lateral swinging of the Rillito Creek at the fan base. Each stream or streamflood deposit and the overlying sheetflood deposit constitute a depositional cycle; deep channels cut in the sheetflood deposits represent an erosional cycle. At times when the Rillito Creek swung toward the base of this fan, erosion was active on the fan, and deep channels were formed. When the river migrated away from the base, these channels were filled by stream and streamflood deposits. Upon accomplishment of graded conditions, sheetfloods may have once more contributed to the building up of this fan.

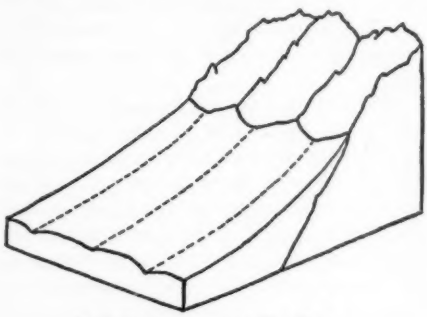


FIGURE 2.—ALLUVIAL FAN WITH TELESCOPE STRUCTURE

As a result of the lateral swinging of a river at the base of a fan a peculiar overall structure may be developed in the fan. It is characterized by younger fans with flatter gradients spreading out from between fan mesas of older fans with steeper gradients; the name *telescope structure* is proposed for it (Fig. 2). Alluvial fans with telescope structure in the Indus basin are illustrated by Drew (1873, p. 454). Conspicuous development of telescope structure is shown in fans of the Santa Catalina Mountains, Arizona, the Black Hills, near Mammoth, Arizona (Pl. 1, fig. 2), and near Indian Garden Springs, Grand Canyon, Arizona.

Effects of Climatic Changes

Of all possible changes in climatic conditions, a variation in the amount of precipitation has the most profound effect on the development of alluvial fans. Increased precipitation results in dissection of fans and the development of a gentler gradient (Eckis, 1928, p. 237; Davis, 1938, p. 1349). In contrast, a decrease in precipitation may be responsible for a period of aggradation and the development of steeper gradients.

Telescope structure may be developed as a result of climatic changes. Individual levels of the telescope structure correspond to arid periods with their tendency for aggradation; the erosion cycles separating aggradation periods mark the humid stages with their tendency for erosion.

The development of telescope structure in alluvial fans at the southern base of the Santa Catalina Mountains, Arizona, is believed to have resulted from climatic changes. These fans formed in Pleistocene and Recent time during which climatic changes were numerous.

Climatic changes may exert an indirect influence on the development of alluvial fans by allowing a through-flowing stream to form at the base of fans. This stream may, in turn, control development of an alluvial fan by changing its base level. Thus, the San Pedro River, north of the Santa Catalina Mountains, Arizona, was a through-flowing stream during a late Pleistocene stage. Alluvial fans extending north from the Santa Catalina Mountains into the San Pedro Valley were dissected by the erosive action of the main river (Davis, 1938).

Effects of Tectonic Movements

Alluvial fans occur commonly at the base of up-faulted mountain blocks. Rejuvenation of tectonic activities along the same fault lines is a frequent phenomenon. Tectonic disturbance is believed the most important reason for changes on alluvial fans, in the opinion of Davis (1938, p. 1348).

If the relief between mountains and fans is increased by tectonic movements, a new stage of deposition may set in; alluvial fans with steeper gradients begin to form. For these newly deposited fans with steeper gradients

the name *superimposed fans* is proposed. Superimposed and original fans do not exhibit distinct boundary lines, and morphologically they cannot be separated. In sections, however, a separation of the two different generations of alluvial fans may be possible (Fig. 3).

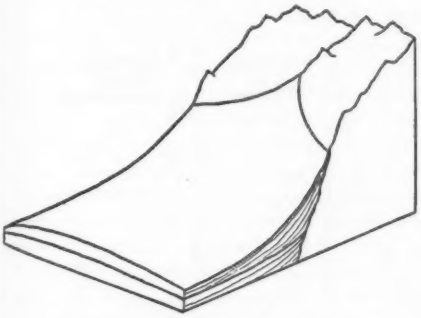


FIGURE 3.—SUPERIMPOSED ALLUVIAL FAN

If the relief between mountains and alluvial fans is decreased by tectonic activities, erosion will be active on the fans, and flatter gradients will be developed. Telescope structure may result. Recognition of rock pediments along the mountain front and their correlation with fan levels may indicate whether telescope structure in fans is due to tectonic movements or to other influences.

In cases where individual parts of an alluvial fan are subjected to tectonic movements, no generalization can be made, and the effects on fans depend on the location, the extent, and the nature of tectonic activities.

Effects of Slumping of Unconsolidated Fan Material

Slumping of unconsolidated fan deposits is a common cause for changes on alluvial fans (Gilbert, 1928, p. 12). An alluvial fan in Nevada on which the effects of slumping were observed is described by Longwell (1930, p. 8).

Slumping of alluvial-fan deposits is favored by the comparatively steep, original dip of strata, the presence of ground water acting as lubricant, the presence of layers with mud-sized constituents acting as sliding planes, and ample voids in coarse-grained layers allowing for considerable settling of strata.

Slumping of unconsolidated fan deposits is responsible for the development of *pseudo-telescope structure*. The difference from true telescope structure may be obvious only in favorable exposures (Fig. 4).

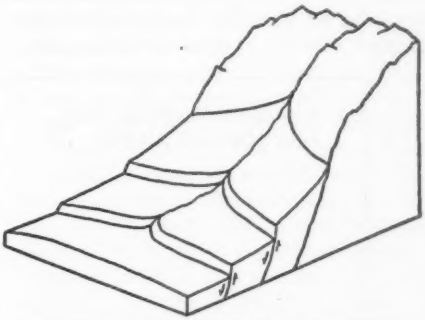


FIGURE 4.—PSEUDO-TELESCOPE STRUCTURE (Modified After Gilbert)

These are the main causes for changes taking place on alluvial fans. Several of these influences may be at work at the same time; their effects may enforce each other, or their tendencies may be opposite and tend to compensate each other. Therefore, the investigator may not always be able to ascribe a particular change in the development of an alluvial fan to any definite cause.

Secondary Alluvial Fans

These form at the base of primary fans. The deposits of secondary alluvial fans are mainly reworked primary fan deposits.

At the base of the large, primary fans at the southern flank of the Santa Catalina Mountains, Arizona, small, secondary fans are in the process of forming (Pl. 1, fig. 3).

GEOLOGY OF ALLUVIAL-FAN DEPOSITS

Facies

The facies of unconsolidated fan deposits are fan gravel, fan sand, and fan mud. The lithified equivalents are fanglomerate (Lawson, 1913, p. 17), fan sandstone, and fan mudstone.

Only one or two of these sedimentary facies normally are developed in an alluvial fan. Many alluvial fans in Arizona, southern Cali-

fornia, and southwestern New Mexico show a unifacial, fanglomeratic development. In contrast, in small alluvial fans of the Black Hills of Arizona, only the sand facies is developed.

If alluvial fans are built up in dry regions, eolian deposits may be incorporated in the fluviatile deposits. In the Aubrey Valley, Arizona, sand dunes rest on the lower parts of some alluvial fans. Further building up of these fans may eventually incorporate the sand dunes within the alluvial-fan deposits (Blissenbach, 1952, p. 123).

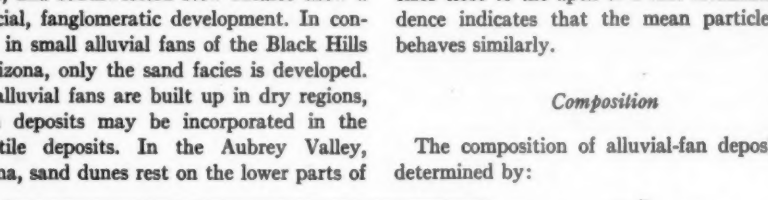


FIGURE 5.—DISTRIBUTION OF MAXIMUM PARTICLE SIZES AND SURFACE ANGLES ALONG A RADIAL PROFILE ON AN ALLUVIAL FAN OF THE SANTA CATALINA MOUNTAINS, ARIZONA

Particle Sizes and Particle-Size Distribution

The particles in alluvial-fan deposits range from boulder to clay size. In general, coarse gravels predominate near the apex of an alluvial fan; material of intermediate particle size may occupy a central zone, and silts and clays in the area close to the base of a fan (Lawson, 1913, p. 330–331; Vaughan, 1922, p. 340; Troxell and others, 1942, p. 322).

Particle-size distribution curves along radial profiles of alluvial fans have been established by the writer for the fans examined. The distribution curve of particle sizes of an alluvial fan of the Santa Catalina Mountains, Arizona, was compiled from a great number of measurements of the maximum particle size (Fig. 5). This distribution curve is typical for alluvial fans; it shows the rapid change in particle sizes close to the apex of a fan. Available evidence indicates that the mean particle size behaves similarly.

Composition

The composition of alluvial-fan deposits is determined by:

(1) the composition of the parent rock from which the fan deposits are derived;

(2) the type and degree of weathering to which the parent rock is subjected;

(3) syngenetic alterations during transport from the source to the site of deposition;

(4) epigenetic alterations effected after deposition.

In arid and semiarid regions, the action of mechanical weathering predominates over the action of chemical weathering. Therefore the composition of the detritus prepared for transport differs only slightly from the composition of the parent rock. In humid regions, however, considerable allowance must be made for alterations effected by the action of chemical weathering.

Syngenetic alterations have little influence on changes in the composition of fan deposits because both the distance of transport and the time during which the detritus is in contact with water are short. Epigenetic alterations are more effective, however. They include the

addition of mineral constituents treated later under Cementation.

Based on the classification of compositional groups proposed by Short, McKee, and Jahns (1953), most alluvial-fan deposits belong to one of two compositional groups, the arkose and the graywacke. Mountains exposing plutonic or coarsely crystalline metamorphic rocks commonly exhibit alluvial fans of arkosic composition. In some places the fan deposits of the gneissic Santa Catalina Mountains, Arizona, show a percentage of feldspars twice as high as the minimum requirement of 20 per cent. Mountains composed of fine-grained metamorphic (mostly volcanic and sedimentary) rocks have alluvial fans with graywacke composition. Fans of the volcanic Tucson Mountains, Arizona, and those of the Aubrey Cliffs, Arizona, exposing sedimentary rocks are of this compositional type.

The composition of fan gravels does not appear to change considerably along a radial profile of a fan. Thus, at the apex of an alluvial fan of the Aubrey Cliffs, Arizona, the composition of fan gravels is 45 per cent sandstones and 55 per cent limestones; at the base of this fan the fan gravels are composed of 48 per cent sandstones and 52 per cent limestones. The slight difference in composition may be accidental, and no definite tendency in compositional change can be recognized.

In some areas fans of contrasting compositional types occur side by side because their apices head in different source rocks. This is illustrated in alluvial fans of the Aubrey Cliffs, Arizona, where large fans are fed from the Kaibab and Toroweap formations and adjacent small fans from the Coconino sandstone.

Sorting

Sorting of alluvial-fan deposits varies widely. In general sorting of alluvial-fan deposits may be regarded as a function of:

(1) the range in particle sizes of the detritus prepared for transport in a mountain area;

(2) the type of transporting and depositing agent;

(3) the distance of transport.

Detritus prepared for transport in a mountain area normally has a wide range in particle sizes. An exception to this trend is found in the Black Hills of Arizona, where the source rock of some alluvial fans is a plutonic rock of granitic texture which weathers to equigranular detritus. The breakdown of this parent rock to uniform particle sizes is responsible for the comparatively high degree of sorting in the resulting fan deposits.

There is considerable variation in the degree of sorting of alluvial-fan deposits depending on whether sheetfloods, streamfloods, or streams are responsible for forming a particular deposit.

Sheetfloods tend to accomplish the least amount of sorting (McGee, 1897, p. 106); Davis, 1938, p. 1344). The coefficient of sorting (Trask, 1932, p. 70) of a sheetflood deposit in a small alluvial fan of the Santa Catalina Mountains, Arizona, is about 3.0. Stream deposits, in contrast, show fair to good sorting. The coefficient of sorting of a stream deposit in a small alluvial fan of the Santa Catalina Mountains, Arizona, is 1.7. Streamflood deposits occupy a position in degree of sorting intermediate between that of sheetflood and of stream deposits.

Most alluvial-fan deposits have been transported only short distances and therefore are not well sorted. However, fairly good sorting of fan deposits may be expected near the base of large fans. Comparatively good sorting is also exhibited in secondary fans due to reworking and selective transport of detritus from primary fans. Thus, deposits of a small, secondary fan at the base of the large, primary Santa Catalina Mountain fans (Pl. 1, fig. 3) show a fair to good degree of sorting.

Roundness

A study of the roundness of fragments in alluvial-fan deposits was carried out on a fan at the southern base of the Santa Catalina Mountains, Arizona. Samples of 25 fragments of about equal size were obtained from 13 points along the radial profile of this fan; determination of the roundness was based on Krumbein's charts (1941, p. 68).

The minimum roundness of fragments in the Santa Catalina Mountain fan studied is 0.2, at the point closest to the apex of this fan; the maximum roundness is 0.7, at the point closest

to the fan base (Fig. 6). The roundness distribution curve (Fig. 6) suggests that the function between the roundness of fan particles and the distance from the apex is approximately linear for this fan.

ing observed in these gravels was attained by long transport in a floodplain environment (Koons, 1948, p. 59). Later faulting brought these gravels into a position from which they could be redeposited as fan gravels. Thus, the

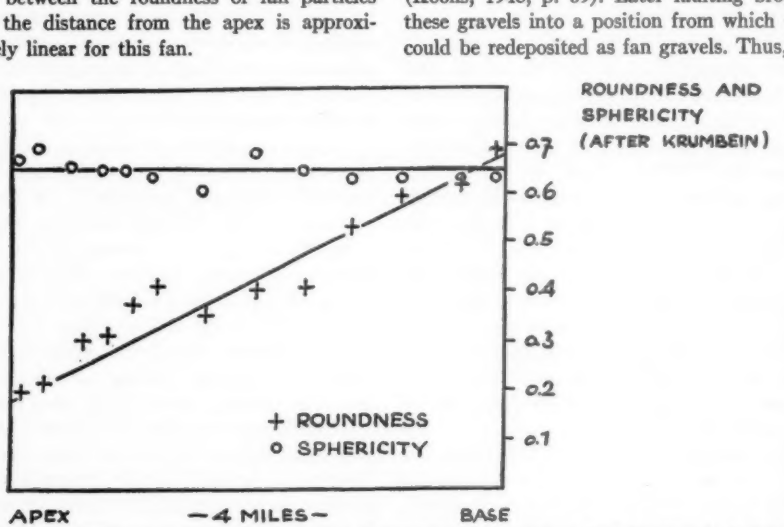


FIGURE 6.—DISTRIBUTION OF ROUNDNESS AND SPHERICITY OF ALLUVIAL FAN PARTICLES ALONG THE SAME RADIAL PROFILE AS FIGURE 5
Base of Santa Catalina Mountains, Arizona

As gneiss particles predominate in alluvial fans of the Santa Catalina Mountains, and as gneiss occupies an intermediate position between the extremely resistant and the highly soluble and friable rocks, a roundness distribution of fan particles in the Santa Catalina Mountain fans may serve as illustration for the normal change in roundness with distance from the apex. Based on the classification of roundness proposed by Short, McKee, and Jahns (1953), fragments of this fan may be described as angular, subangular, and subrounded. As the radial extent of alluvial fans may greatly exceed the 4-mile extent of the Santa Catalina Mountain fans, rounded fragments must be expected on larger fans. The common conception of alluvial-fan deposits being necessarily characterized by angularity of their fragments (Chawner, 1935, p. 259) must therefore be abandoned.

Pliocene gravel deposits at Blue Mountain, northwest of Seligman, Arizona, are described in a recent study by the writer (1952, p. 120). The Blue Mountain gravels were deposited in part on alluvial fans. The high degree of rounding in these fan gravels was not accomplished on an alluvial fan but in a floodplain environment.

Sphericity

Sphericity meansurements in alluvial-fan deposits of the Santa Catalina Mountains, Arizona, were made of the samples obtained for the determination of roundness. For the determination of the sphericity of fan particles, Krumbein's charts (1941, p. 66) were employed. The mean sphericity of particles in this fan is 0.65. The maximum and the minimum values of the sphericity of fan particles along the 4-mile extent of this fan deviate but 7 per cent from the value for the mean sphericity (Fig. 6).

No major change in the sphericity of fan particles is expected on any particular fan as long as there is no change in composition or facies of the fan deposits.

Matrix

Particles forming the matrix of alluvial-fan gravels range from sand to mud size. Finer

particles predominate in the matrix of sheetflood deposits, coarser particles in that of stream deposits. A typical sample from the matrix of a sheetflood deposit of a small alluvial fan of the Santa Catalina Mountains, Arizona, shows a mean particle size of 0.15 mm. The mean particle size of the matrix of a stream deposit in the same alluvial fan is 0.7 mm.

The composition of the matrix in a fan deposit determines its compositional classification. Commonly the matrix is arkosic or graywacke.

A secondary matrix is a common feature in fan gravels. This results from removal of an original interstitial fill between fan gravels and later filling of interstices. This process may be seen in action on alluvial fans of the Santa Catalina Mountains, Arizona.

Cement

Calcium carbonate is the commonest cement in alluvial-fan deposits in most areas. It is precipitated from ascending or descending ground water and coats fan particles as solid layers or concretions, or is disseminated as minute calcite crystals in the matrix. If all available pore space is filled with calcium carbonate the cement may constitute up to 30 per cent of the volume of the composite fan deposit. Limonite cement is encountered in a few places in alluvial-fan deposits.

Color

The colors of alluvial-fan deposits are due largely to the types of rock of which they are composed, but they vary somewhat with the climate.

Mudflow deposits of alluvial fans of the Santa Catalina Mountains, Arizona, are light yellow; the interbedded stream deposits are a darker yellow. Where calcium carbonate cement is abundant, alluvial-fan deposits range from gray to white.

Porosity

The porosity in pebble lenses with loose packing, observed in a small, recent fan of the Santa Catalina Mountains, Arizona, was estimated to be 30 per cent. Owing to cementation the porosity in the deposits of older fans along these mountains does not exceed 15 per cent. The porosity of alluvial-fan deposits may attain any value between the above extremes and zero, as controlled by sorting, degree of packing, roundness of particles, compaction, and cementation.

In general poorly sorted mudflow deposits have low porosity; well-sorted stream deposits have high porosity.

Permeability

The permeability of alluvial-fan deposits depends on their porosity and, in addition, is a function of the size of interstitial space. In addition to their low porosity, mudflow deposits have small interstitial openings. Molecular forces therefore tend to retain water, and ground water cannot circulate freely. Thus, mudflow deposits are ideal aquicludes. In contrast, stream deposits in alluvial fans have comparatively large pore spaces which account for the ease with which ground water can move through them. Stream deposits therefore are good aquifers.

Alluvial fans built up by streams and mudflows are, in terms of ground-water geology, "alternating aquifers and aquicludes". This property together with the original dip of the fan strata tends to make an alluvial fan an ideal site for the recovery of artesian water (Tolman, 1937, p. 364–365; Eardley, 1938, p. 1408).

Sedimentary Structures

Alluvial-fan deposits are laid down in beds approximately parallel to the surface of a fan. Thus, the surface angles of alluvial fans are repeated within the fan strata. The classification adopted for the angles of dip of a fan surface may also be applied to the dip of the strata in an alluvial fan; the dip may be described as steep, gentle, or flat.

The overall stratification of most alluvial fans is fair. In fans built up through their entire extent by mudflows, stratification may be poor. In general mudflow deposits form one solid layer. Mudflow deposits show distinct contacts with stream deposits (Fig. 7). Deposits laid down by streams or moderate streamfloods show good stratification; they are sub-

divided into thinner layers (Pl. 2, fig. 2). Where the mean particle size of these deposits allows such development, lamination is common.

Cross-stratification is well developed in deposits formed by streams on alluvial fans. Current ripple marks may be present (Twenhofel, 1950, p. 71). The most common and conspicuous type of cross-stratification is due to the filling of channels that have been cut into the underlying deposits (Fig. 7; Pl. 2, fig. 2).

The thickness of individual strata in fans ranges from a fraction of an inch to 15 or 20 feet. Most deposits laid down by streams on alluvial fans are in layers from 1 inch to 1 foot in thickness. Successive stream deposits may attain a vertical extent of several tens of feet.

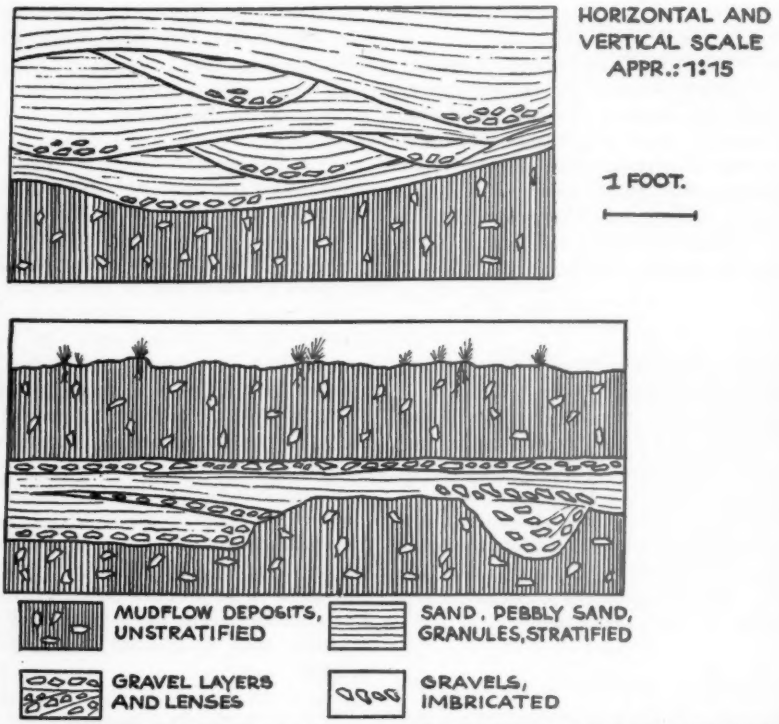


FIGURE 7.—CHARACTERISTIC SEDIMENTARY STRUCTURES IN AN ALLUVIAL FAN OF THE SANTA CATALINA MOUNTAINS, ARIZONA
Main direction of transport from right to left.

Individual mudflow deposits range in thickness from 1 foot on small fans to 15 or 20 feet on large fans. Alternating mudflow deposits and series of stream deposits commonly are of comparable thickness.

Particles in mudflow deposits frequently stand on edge, and in all other normal positions

PLATE 1.—ALLUVIAL FANS

FIGURE 1.—COMPOUND ALLUVIAL FAN ON THE SOUTH FLANK OF THE SANTA CATALINA MOUNTAINS, ARIZONA

FIGURE 2.—SMALL ALLUVIAL FAN WITH TELESCOPE STRUCTURE, BLACK HILLS, ARIZONA

FIGURE 3.—SECONDARY ALLUVIAL FANS (S) AT THE BASE OF PRIMARY FANS (P), SANTA CATALINA MOUNTAINS, ARIZONA

FIGURE 1

FIGURE 2

FIGURE 3

ALLUVIAL FANS

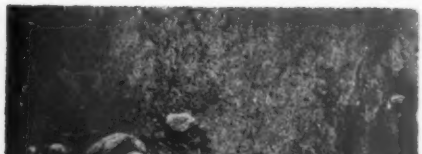


FIGURE 1

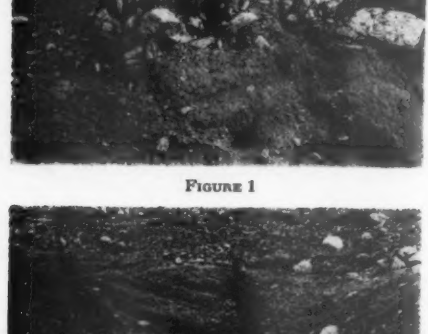


FIGURE 2

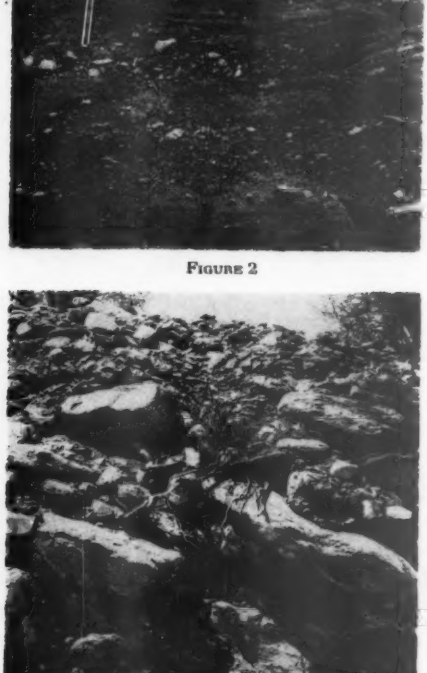


FIGURE 3

ALLUVIAL-FAN DEPOSITS

of repose (Fig. 7; Pl. 2, fig. 1) they do not exhibit any preferred orientation. Imbrication is pronounced in deposits laid down by streams or moderate streamfloods wherever the shape of the particles allows such development. The gneissic bedrock of the Santa Catalina Mountains, Arizona, tends to break down to platy or disc-shaped fragments; therefore imbrication is distinct in certain fan deposits at the base of these mountains (Fig. 7). The disc-shaped fragments are inclined toward the site of derivation as might be expected within a fluviatile environment. The angle of inclination reaches a maximum of 30° among particles of pebble and cobble size; boulders, 100 to 200 cm in diameter, have no inclination greater than 25° (Pl. 2, fig. 3).

Stream and streamflood deposits are rudely lenticular. They are deposited in channels and follow in their extent the winding courses of the former channels. Sheetflood deposits, in contrast, are laid down in definite sheets. They terminate, however, with sharp, abrupt edges (Chawner, 1935, p. 256).

Organic Contents

Fossils are rare in the deposits of an alluvial fan, but there may be local occurrences of abundant fossils due to sudden floods which caused wholesale killing and rapid burial (Twenhofel, 1950, p. 71). Fossil animals were not found within the alluvial fans studied by the writer. Fossil plants or plant fragments, however, appear to be abundant in certain fan deposits.

Ancient Alluvial-Fan Deposits

Occurrence

Numerous occurrences of ancient alluvial-fan deposits are listed by Twenhofel (1950, p. 72). According to him, the earliest recorded alluvial-fan deposits are represented by some of the coarse sediments of the Huronian and some of the Keweenawan conglomerates and sandstones of the Lake Superior region. Alluvial-fan deposits have been recognized in many younger sedimentary formations in all parts of the world.

Relative Abundance

The failure to recognize alluvial-fan deposits in the geologic column to the extent they are known among Recent sediments may be explained by the following factors:

(1) The bold relief necessary for the development of alluvial fans was not present in many parts of the geologic past (Barrell, 1925, p. 292).

(2) A combination of bold relief and aridity favorable for the building of alluvial fans was not common in the geologic past (Lawson, 1913, p. 334).

(3) The dominance of land, as in the Quaternary, is exceptional in geologic history (Lawson, 1913, p. 334).

(4) The process of peneplanation eroding the mountains to their roots also affects the alluvial fans adjacent to the mountains and tends to remove them. Only where some unusual feature such as extensive downfaulting, like that involving the Newark series, has protected the alluvial-fan deposits from further erosion may they be preserved.

Alluvial-fan deposits have not always been described as such but have been classified as angular conglomerates or sedimentary breccias (Lawson, 1913, p. 333). In the opinion of the present writer, the misconception common in geologic literature that fan deposits are necessarily unstratified, composed of highly angular particles, poorly sorted, and without distinctive

---

Plate 2.—ALLUVIAL-FAN DEPOSITS

Figure 1.—Fanglomerate Deposited by Mudflow

Poor sorting, random orientationof particles. Largest particles in picture are 10 centimeters in diameter. Santa Catalina Mountain fans, Arizona

Figure 2.—Fanglomerate Deposited by Stream or Moderate Stream-Flood

The effect of channel cut-and-fill is pronounced. Ruler is 15 centimeters. Main direction of deposition from left to right. Santa Catalina Mountain fans, Arizona.

Figure 3.—Imbrication in Large Particles

Up to 15° centimeters. Direction of transport from right to left. Close to the apex of an alluvial fan of the Santa Catalina Mountains, Arizona

sedimentary structures may in part be responsible for such classification.

Criteria for Recognition of Ancient Alluvial-Fan Deposits

The following criteria can be applied where erosion or tectonic movements have obscured the original surface of a deposit. The recognition of an ancient alluvial-fan deposit must rest on a number of criteria.

The facies of alluvial-fan deposits is coarse detrital as a rule but may be fine detrital under exceptional conditions. Individual particles in these deposits may range from boulder to clay size. A rapid change in the maximum or the mean particle size suggests an origin on an alluvial fan. The presence of sheetflood deposits makes such an assumption more probable. Arkoses and graywackes are the compositional groups in which alluvial-fan deposits commonly range.

The coefficient of sorting (Trask, 1932, p. 70) of fan deposits may range from about 1.5 to 3.0. Fragments in alluvial-fan deposits are angular, subangular, subrounded, and rounded. The change in roundness is rapid close to the area of derivation.

The matrix of fan gravels is sand or mud in size; arkose or graywacke in composition; of primary or secondary origin. The original dip of fan strata may be steep, gentle, or flat. Alluvial-fan deposits have good to poor stratification depending on the relative abundance of unstratified mudflow deposits and stratified stream deposits. Channel cut-and-fill is common in the deposits formed by streams and moderate streamfloods. In these sediments imbrication is pronounced; particles are inclined toward the site of derivation; the inclination commonly does not exceed 30°.

The thickness of individual fan strata ranges from a fraction of an inch to 15 or 20 feet. Stream and streamflood deposits follow the winding courses of former channels; sheetflood deposits form definite sheets.

In some places the fanhead area may be in contact with a talus slope, and in these places fan deposits may be expected to interfinger with talus-slope deposits. The following characteristics of talus-slope deposits may help to separate these sediments from the strata of an alluvial fan. Talus-slope deposits are almost devoid of sedimentary structures as deposition takes place by gravitational sliding. Particles are larger than those found in the adjacent fan deposits. Roundness of talus-slope particles is expected to be of a low degree. Platy fragments may be arranged parallel to the former surface of a talus slope whose original dip may range from 10° to 30°.

At the base of a fan the deposits of an alluvial fan may interfinger with the sediments of a floodplain or a playa lake. The deposits of a playa lake are easily distinguished from the fan deposits as they comprise sediments of clay size and chemical deposits both of which are uncommon or absent in alluvial-fan strata. Floodplain deposits have better stratification than those of alluvial fans; mudflow deposits are uncommon in a floodplain environment. Sorting of the deposits and rounding of sedimentary particles are likely to attain better values in a floodplain than on an alluvial fan. Floodplain deposits are laid down at angles commonly lower than 1°.

This discussion offers some suggestions for the recognition of the climatic conditions under which an ancient fan deposit was formed. With increased precipitation, mudflow deposits become progressively less numerous. From findings on recent alluvial fans it appears that no mudflow deposits are expected in alluvial-fan strata if the mean annual precipitation exceeds 20 or 25 inches. If mudflow deposits make up 50 per cent of an alluvial fan, 5 to 15 inches of annual rainfall appears to be a conservative estimate for the climate under which these fans were formed. The types of rainfall and their distribution through the year as well as the relative abundance of vegetation are factors which control the development of certain types of deposits, in addition to the total amount of precipitation.

Abundance of mudflow deposits in an alluvial fan does not necessarily imply the forming of this fan in a tropical or hot, arid region; it may have been formed in zones with moderate temperatures or even in periglacial environment. Conspicuous development of mudflow deposits may be expected especially along the margins of glaciated areas where they is no movement of water during the colder season

and a sudden release of abundant water with the beginning of a warmer season. Fragments in these deposits probably bear ample evidence in their surface texture as to their source within the glaciated area.

Humid conditions during the formation of an ancient alluvial fan may be assumed if the composition of the deposits of the fan is considerably different from that of the parent rock indicating a predominance of chemical over mechanical weathering.

References Cited

Barrell, J., 1925, Marine and terrestrial conglomerates: Geol. Soc. Am. Bull., v. 36, p. 279–342.

Blackwelder, E., 1931, Desert plains: Jour. Geology, v. 39, p. 133–140.

Blissenbach, Erich, 1952, Geology of the Aubrey Valley, south of the Hualpai Indian Reservation, Northwest Arizona: Plateau, v. 24, no. 4, p. 119–127.

Buwalda, J. F., 1951, Transportation of coarse material on alluvial fans (Abstract): Geol. Soc. America Bull., v. 62, p. 1491.

Chawner, W. D., 1935, Alluvial fan flooding: The Montrose, California, flood of 1934: Geog. Rev., v. 25, p. 255–263.

Dana, J. D., 1894, Manual of geology: New York, 4th ed., 1087 pages.

Davis, W. M., 1898, Physical geography: Boston, 428 pages.

—— 1938, Sheetfloods and streamfloods: Geol. Soc. America Bull., v. 49, p. 1337–1416.

Drew, F., 1873, Alluvial and lacustrine deposits and glacial records of the upper Indus basin: Geol. Soc. London Quart. Jour., v. 29, p. 441–471.

Eardley, A. J., 1938, Sediments of Great Salt Lake, Utah: Am. Assoc. Petroleum Geologists Bull., v. 22, p. 1305–1411.

Eckis, R., 1928, Alluvial fans in the Cucamonga District, Southern California: Jour. Geology, v. 36, p. 224–247.

Gilbert, G. K., 1928, Studies of the Basin-Range structure: U. S. Geol. Survey Prof. Paper 153, 92 pages.

Grabau, A. W., 1913, Principles of stratigraphy: New York, A. G. Seiler Co., 1185 pages.

Koons, D., 1948, Geology of the eastern Hualpai Reservation: Plateau, v. 20, no. 4, p. 53–60.

Krumbein, W. C., 1937, Sediments and exponential curves: Jour. Geology, v. 45, no. 6, p. 577–601.

—— 1941, Measurement and geologic significance of shape and roundness of sedimentary particles: Jour. Sed. Petrology, v. 11, no. 2, p. 64–72.

Lahee, F. H., 1941, Field geology: 4th ed., McGraw-Hill Book Co., New York, 853 pages.

Lawson, A. C., 1906, The geomorphogeny of the Techachapi Valley System: Univ. Calif. Pub., Dept. Geology, Bull., v. 4, p. 431–462.

—— 1913, The petrographic designation of alluvial fan formations: Univ. Calif. Pub., Dept. Geology, Bull., v. 7, p. 325–334.

—— 1915, The epigene profiles of the desert: Univ. Calif. Pub., Dept. Geology, Bull., v. 9, p. 23–48.

Longwell, C. R., 1930, Faulted fans west of the Sheep Range, South Nevada: Am. Jour. Sci., 5th ser., v. 20, p. 1–13.

Longwell, C. R., Knopf, A., and Flint, R. F., 1948, Physical geology: 3 rd ed, New York, John Wiley and Sons, 602 pages.

McGee, W. J., 1897, Sheetflood erosion: Geol. Soc. America Bull., v. 8, p. 87–112.

Miller, W. J., 1926, Geology: New York, D. Van Nostrand Co., 555 pages.

Pack, F. J., 1923, Torrential potential of desert waters: Pan-Am. Geologist, v. 40, p. 349–356.

Scott, W. B., 1932, An introduction to geology, vol. 1: New York, Macmillan Co., 604 pages.

Short, M. N., McKee, E. D., and Jahns, Richard, 1953, Hand specimen petrology: New York, John Wiley and Sons, (In press).

Tolman, C. T., 1937, Groundwater: 1st ed. New York, McGraw-Hill Book Co., 593 pages.

Trask, P. D., 1932, Origin and environment of source sediments of petroleum: Gulf Pub. Co., Houston, Texas, 323 pages.

Troxell, H. C., *et al.*, 1942, Floods of March 1938 in southern California: U. S. Geol. Survey Water-Supply Paper 844, 399 pages.

Twenhofel, W. H., 1950, Principles of sedimentation: 2d ed., New York, McGraw-Hill Book Co., 673 pages.

Vaughan, F. E., 1922, Geology of the San Bernardino Mountains north of San Gorgonio Pass, California: Calif. Univ., Dept. Geol. Sci., Bull., v. 13, no. 9, p. 319–411.

Landshut/Bayern, Lebühlstrasse 2, Germany

Manuscript Received by the Secretary of the Society, October 29, 1952

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 65, PP. 191–194, 1 PL. FEBRUARY 1954

Short Notes

DEEP-SEA CHANNELS AND DELTA OF THE MONTEREY SUBMARINE CANYON

BY ROBERT F. DILL, ROBERT S. DIETZ, AND HARRIS STEWART

Sonic and spot soundings by the U. S. Coast and Geodetic Survey and subsequent contouring of these data by Shepard and Emery (1941) and the U. S. Coast and Geodetic Survey (Charts 5302 and 5402) show that the continental terrace off Monterey Bay, California, is incised by one of the world's largest submarine canyons, known as the Monterey Submarine Canyon. However, because of the sparseness of soundings at the base of the continental slope, the nature of the region around the canyon mouth remained practically unknown until recently.

That the Monterey Canyon might have a channel extending out over the deep-sea floor like that of the Hudson Canyon (Heezen Ewing, and Ericson, 1951) was first suspected in 1949 when the fathogram from a cruise to the Arctic showed a small notch in the deep-sea floor many miles seaward of the Monterey proper. More recently, in connection with SOFAR operations in the Monterey area, two more crossings of this deep-sea channellike feature were made. Then, in December 1951, several sounding lines were run with the Navy Electronics Laboratory vessel, the EPCE(R) 857—a 180-foot converted escort patrol craft equipped with an NMC-2 echo sounder. This survey furnished sufficient data to establish the presence of a deep-sea extension of Monterey Canyon. A deltalike sedimentary apron extending out from the mouth of the canyon was also revealed. The survey showed that the channels on the apron were in places bordered by leveelike banks. Two short cores were taken—one in the axis of the channel and the other from a "natural levee."

Plate 1 was made from the recently obtained fathograms plus soundings from the United States Coast and Geodetic Survey. The contours of the shelf down to 1700 fathoms are well established by numerous soundings so they were taken directly from the U. S. Coast and Geodetic Survey Chart 5402 (contour interval 100 fathoms). Comparatively few soundings previously were available from the deep-sea floor so it is not surprising that the interpretation as shown in Plate 1 is quite different from that shown on chart 5402 and on the Shepard and Emery (1941) chart. This new interpretation shows a smooth deltalike fan extending out from the base of the continental slope. Two large channels and possibly a smaller one cut into this delta. The delta cannot be identified below the 2100-fathom contour. If this is the limit of the depth, the delta is about 60 miles long and 60 miles wide at its maximum cross section. It extends southwest from the mouth of the Monterey Canyon with a regional gradient of 60 feet per mile. If the delta has built up from a base level of 2100 fathoms, it has a total volume of about 26 cubic miles of sediment. By comparison the volume of Monterey Canyon is about 53 cubic miles.

The Monterey Canyon proper extends seaward for 50 miles from near the shore line in Monterey Bay to the base of the continental slope at 1700 fathoms. The topography of the canyon has been described in considerable detail by Shepard and Emery (1941, p. 72–78), Woodford (1951), and Crowell (1952, p. 75). In the 20 miles between the 1700- and 1800-fathom contours the canyon has a shallow U-shaped cross section. The gradient of the axis is low (30 feet per mile), so that the can-

yon appears to lose its identity as a valley form. However, a short distance beyond the 1800-fathom contour, a channel again notches the deep-sea floor. In the vicinity of the 1850-fathom contour, two "distributary" channels break away from the main channel. All three channels assume courses roughly down the regional gradient of the delta. The west or main channel has a maximum width of 2.8 miles and a maximum depth of 750 feet. Its length is unknown, but it has been traced for 40 miles. A leveelike ridge with a relief of 90 feet is present on the outside of the west channel where it turns south. The central channel has a maximum width of 1.2 miles, a maximum depth of 480 feet, and an established length of 20 miles. However, where traverses were made, it has no leveelike ridges. It is also slightly shallower than the west channel. The east channel is small; its presence is based on only one crossing, so its extent is unknown.

Two short cores were obtained with a small gravity corer. The core 47 cm in length was taken from the axis of the main channel; the other, obtained from the leveelike ridge of the main channel, was 43 cm long. The core from the channel showed stratification and consisted of alternating layers of very fine-grained sand and a green micaceous terrigenous mud. The boundaries between the layers were sharp, but there was no evidence of graded bedding within the layers. The sand layers in this channel core were composed predominantly of a well-sorted quartz sand and ranged in thickness from 0.25 to 0.50 inch. The majority of the individual grains were angular to subangular with no evidence of frosting or other surface markings. The heavy-mineral suite indicates derivation from metamorphic and volcanic rocks of continental origin (E. L. Hamilton, Personal communication). The absence of any Foraminifera in the sand layers made it impossible to determine whether the sands had been transported from shallower water. However, this lack of Foraminifera does indicate rapid deposition.

The fine-grained material between the sand layers consisted of homogeneous green terrigenous mud in layers 0.25 to 7 inches thick. The only material larger than silt size in these mud layers was mica and a few grains of presumed glauconite. The mud layers contain abundant Foraminifera. According to O. L. Bandy (Personal communication) the fauna is dominated by *Uvigerina senticosa*—a deep-water species typical of depths between 8000 to 12,000 feet (1300–2000 fathoms). No forms typical of shallow water were found.

The core taken from the levee bank consisted of homogeneous green micaceous terrigenous mud. This material is strikingly similar to the mud layers of the other core. Markings made by burrowing animals were present throughout the core.

The writers believe that subaerial processes did not form the delta and the channels at the base of the Monterey Canyon because: (1) a relative change of sea level of approximately 2 miles seems improbable, (2) deep-sea channels formed subaerially would probably have been destroyed in the passage of the strand line across them during a relative change in sea level of 2 miles, (3) the channels are well adjusted to the present slope of the sea floor; this would be unlikely if the channels are ancient subaerial channels that have been tectonically downdropped.

The processes that form submarine deltas can best be interpreted by the study of the surface features of this delta. One of the most important features is the leveed channels that meander across the broad submarine delta-fan much as do the subaerial leveed channels of a river. The similarity between the subaerial delta distributaries and the deep-sea channels is so striking that probably they are formed by similar processes. Since both subaerial and sublacustrine sediment-charged currents can form channels with leveed banks, it is reasonable to assume that deep-sea leveed channels are caused by sediment-charged currents. Menard and Ludwick (1951) have suggested that turbidity currents might produce leveed submarine channels. Although their existence on the sea floor has never been definitely demonstrated, such currents appear to offer the most logical explanation for these deep-sea channels.

The characteristics of the sediment collected neither strongly confirm nor deny the turbidity-currents hypothesis. The two short cores certainly are not an adequate sampling of the

environment. However, in conformity with the principle of isostasy, the bedrock in the abyssal sea-floor region of the channels is presumably dense basic rock. If so, the acidic mineral assemblage shows that this sediment was derived from the continental block. Although the writers believe that much of this sediment was probably carried out by turbidity currents, other processes can also account for this mineral assemblage. The presence of fine sand layers in the core from the channel and their absence in the core from the levee suggests that the currents which deposited the sand layer may also have formed the channels. Inasmuch as only two cores were collected, the presence of sand layers in one core and not in the other may be purely fortuitous. The sand layers also indicate some type of bottom current, and the absence of an indigenous Foraminifera fauna in this sand suggests a rapid emplacement of these sand layers.

The delta and the deep-sea channels may have developed as follows: Sedimentation causes oversteepening and eventual slumping of sediment in Monterey Canyon, which initiates a turbidity current. Because of the steep gradient, these currents accelerate as they flow down the axis of the canyon, probably increasing in size by picking up sediment. Once a flow reaches the deep-sea floor the low gradient of the bottom reduces the velocity, so that the flow becomes predominantly depositional. In this way a wedge of sediments is built up to form an undersea alluvial fan the surface of which is a plane of equilibrium. The fan slowly fills the mouth of the original V-shaped canyon as it laps up on the continental slope. As successive flows leave the canyon mouth, friction against its moving front and lateral mixing with the surrounding water cause a decrease in the velocity at the edges of the flow resulting in deposition along the sides. Subsequent flows tend to be channeled between these lateral deposits or levees, building them up and extending them seaward. Thus the channels are developed by a process of erosion and differential deposition along a profile of equilibrium. The number of channels and the shape of the delta proper indicate that these channels may meander across the surface as do the channels of land deltas, spilling their sediments first in one area and then in another so that the end result is a large fan-shaped delta similar to those found on the continents.

References Cited

Crowell, J. C., 1952, Submarine canyons bordering central and southern California: Jour. Geology, v. 60, p. 58–83.

Heezen, B. C., Ewing, M., and Ericson, D. B., 1951, Submarine topography in the North Atlantic: Geol. Soc. America Bull., v. 62, p. 1407–1409.

Menard, H. W., and Ludwick, J. C., 1951, Applications of hydraulics to the study of marine turbidity currents: Soc. Econ. Mineralogists and Paleontologists Spec. Pub. no. 2, p. 2–13.

Shepard, F. P., and Emery, K. O., 1941, Submarine topography off the California coast: Canyons and tectonic interpretation: Geol. Soc. America Spec. Papers, no. 31, 171 p.

Woodford, A. O., 1951, Stream gradients and Monterey Sea Valley: Geol. Soc. America Bull., v. 62, p. 799–852.

Navy Electronics Laboratory, San Diego 52, California; Navy Electronics Laboratory, San Diego 52, California; Scripps Institution of Oceanography, La Jolla, California

U. S. Navy Electronics Laboratory Professional Contribution No. 11

Scripps Institution of Oceanography Contribution No. 669

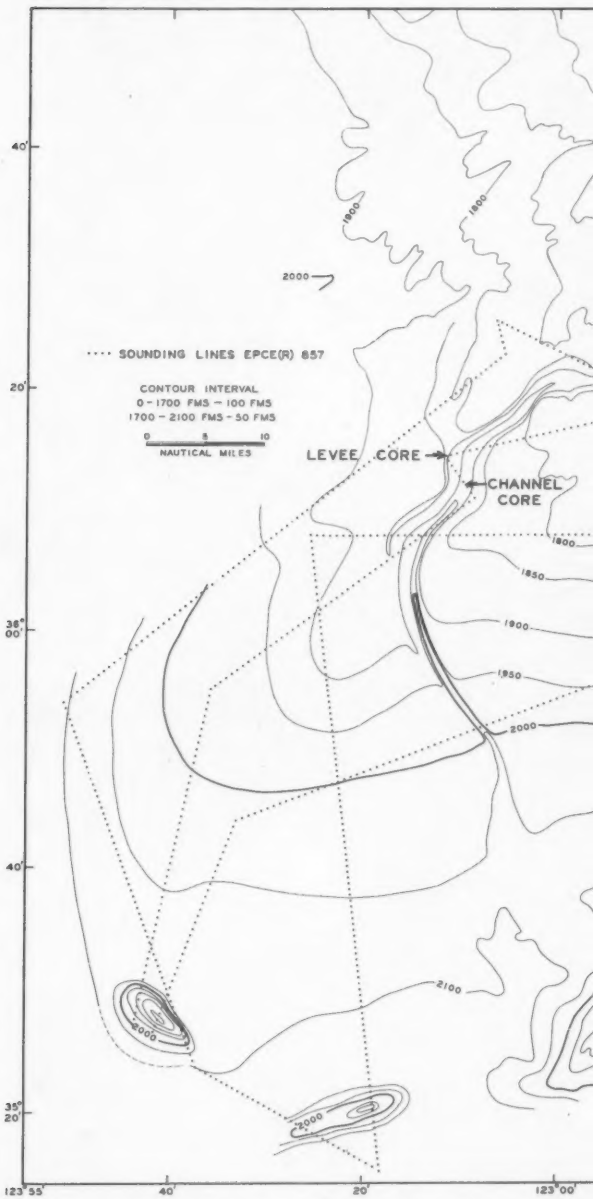
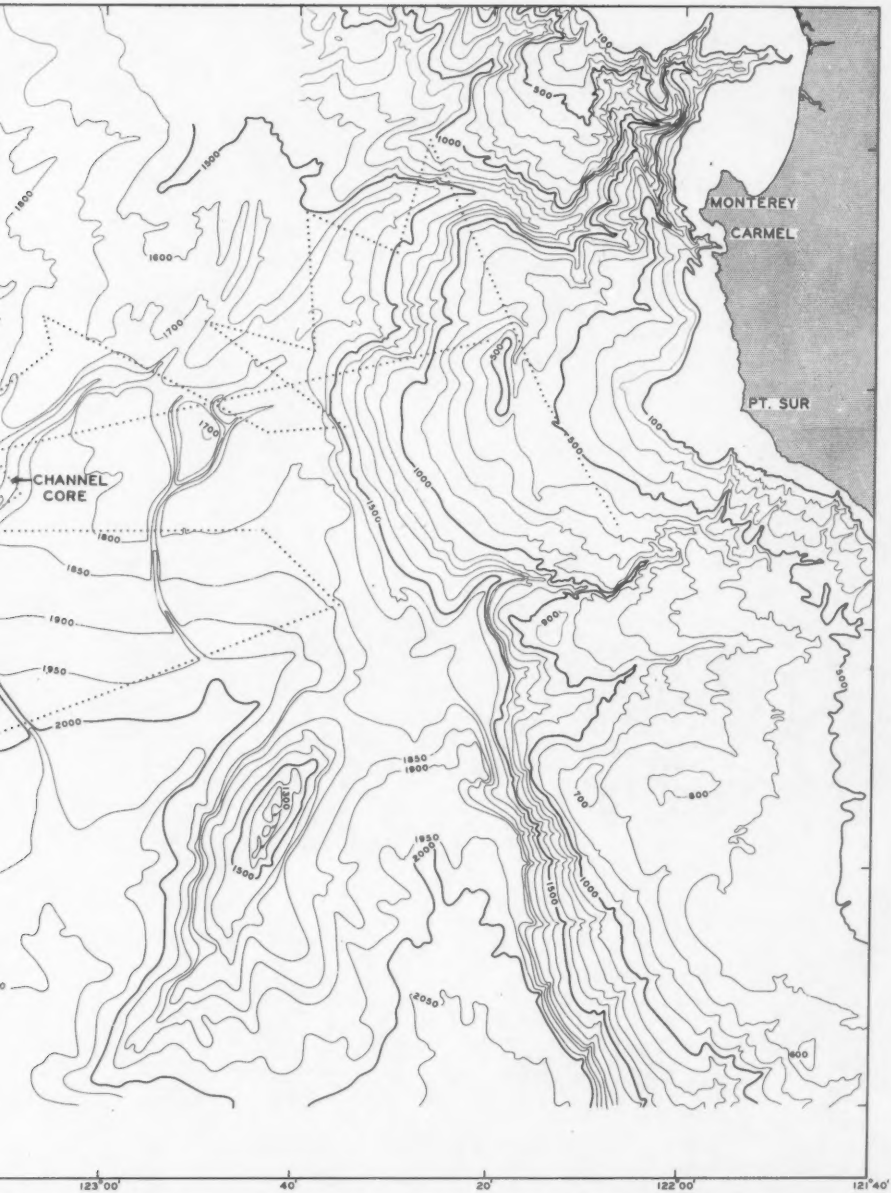


CHART SHOWING THE DEEP-SEA CHANNELS

Note that soundings less than 1800 fathoms are contour
50-fathom interval. The new sounding lines, on which thi
shown by the lines of dots. The other contours outside th
from Shepard and Emery (1941).



CHANNELS AND DELTA OFF THE MONTEREY SUBMARINE CANYON

ns are contoured with a 100-fathom interval and those deeper than 1800 fathoms with a
, on which this interpretation of the topography off the Monterey canyon is based, are
urs outside this area are taken from the U. S. Coast and Geodetic Survey chart 5402 or

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 65, PP. 195-200, 2 FIGS. FEBRUARY 1954

GRAVITY ANOMALIES AND STRUCTURE OF THE WEST INDIES PART II

BY J. LAMAR WORZEL AND MAURICE EWING

From new seismic data and the gravity anomalies, the structure section north of Puerto Rico was deduced by Ewing and Worzel (1954, p. 166). The data were very sparse, and the structure section derived was therefore quite general. It was possible however to show that a large downward deflection of the crust into a tectogene did not occur, but that the large negative gravity anomaly was due to a small downward deflection of the crust (about 7 km) and a large accumulation of sediments. It was suggested that this is the first stage in the formation of an island arc.

According to this suggestion an island such as Puerto Rico was formed as follows: a trench formed, filled rapidly with sediments swept from ocean areas and from near-by island arcs or land largely through the action of turbidity currents, the sediments were granitized, and the basaltic layer beneath the trench was contaminated producing andesitic lavas (the water accompanying the andesitic volcanic activity being derived from sea water associated with the sediments). This process implies that trenches, which habitually form in oceanic crust near the edges of continents, upon relaxation or reversal of the force responsible for the trench, collect most of the continental and oceanic debris forming an island arc. The area between island arc and continent is eventually added to the continent.

In the spring of 1953 measurements on board USS DIABLO (SS479) added 44 new gravity stations to a profile extending from about 200 miles due north to 200 miles due south of San Juan, Puerto Rico. Figure 1 shows the location of these stations as well as the locations of the gravity stations and seismic stations used in the earlier work. The Columbia University Frost type geodetic gravimeter was taken along, and one observer was put ashore on Puerto Rico to make additional land measurements. Four hundred new gravity stations were made in a network covering the whole island of Puerto Rico. These data are soon to be reported by Mr. Lynn Shurbet.

Figure 2A shows the known seismic and topographic data along the line of profile. Figure 2B shows the layers and densities from which a computed gravity curve was obtained. Figure 2C shows the gravity data and the computed curves. The solid dots are the gravity data used in Part I. The open circles show the free-air anomalies at the new gravity-at-sea stations. The dotted curve over the island of Puerto Rico shows the modified Bouguer anomalies along the line of profile. The double dash-and-dot curve over the island of Puerto Rico shows the average modified Bouguer anomalies of five profiles across the island of Puerto Rico made available by Mr. Shurbet prior to publication. The solid line is the anomaly curve computed from the mass distribution used in Part I. The dashed curve shows the anomaly curve computed for the mass distribution shown in Figure 2B, on the assumption that measurements are made at sea level. This computed anomaly curve quite satisfactorily fits the free-air anomalies for sea stations and the modified Bouguer anomalies for land stations. The oscillations in the computed curve would be smoothed out if the computations were made for the blocks having inclined boundaries instead of the rectangular forms used here. In this form the computations would have been more laborious and would have added little to the picture.

Figure 2D shows the section derived from the seismic and the gravity data without vertical exaggeration. The section to the north of San Juan is changed from our previous work by the addition of more detail, which is now justified by the additional data. South of Puerto Rico the new gravity minimum of almost −100 milligals is accounted for by a slight downward deflection of the crust and a thick column of sediments. Because of lack of time, the gravity observations at sea could not be continued in close to the south shore of Puerto Rico.

This more detailed and extended profile confirms all of the interpretations and conclusions made in our previous report and requires no modifications of the hypothesis and the implications introduced there.

Thanks are due Mr. Lynn Shurbet and Mr. Hugh Traphagen who made all the new gravity observations. Mrs. Elizabeth S. Skinner did most of the calculations for the gravity-at-sea data, and Mr. Lynn Shurbet did all the calculations of the land gravity measurements. Miss Annette Trefzer did the drafting.

To the officers and men of USS DIABLO we wish to express our appreciation for their interest and assistance in making the gravity observations at sea.

Professor Raoul C. Mitchell and Don Luis Stefani of the University of Puerto Rico provided much valuable advice and assistance in the course of the measurements on Puerto Rico.

Transportation was generously provided by Dr. R. Fernandez Garcia, Director, Industrial Research Department, Economic Development Administration of the Commonwealth of Puerto Rico.

Dr. Raymond J. Smith of the United States Geological Survey provided much assistance.

Most of this work was carried out under contract N6 onr 271 with the office of Naval Research of the Department of the Navy. Financial assistance for the land gravity work was made available by the University of Puerto Rico.

LAMONT GEOLOGICAL OBSERVATORY (COLUMBIA UNIVERSITY), PALISADES, N. Y.
LAMONT GEOLOGICAL OBSERVATORY CONTRIBUTION NO. 102

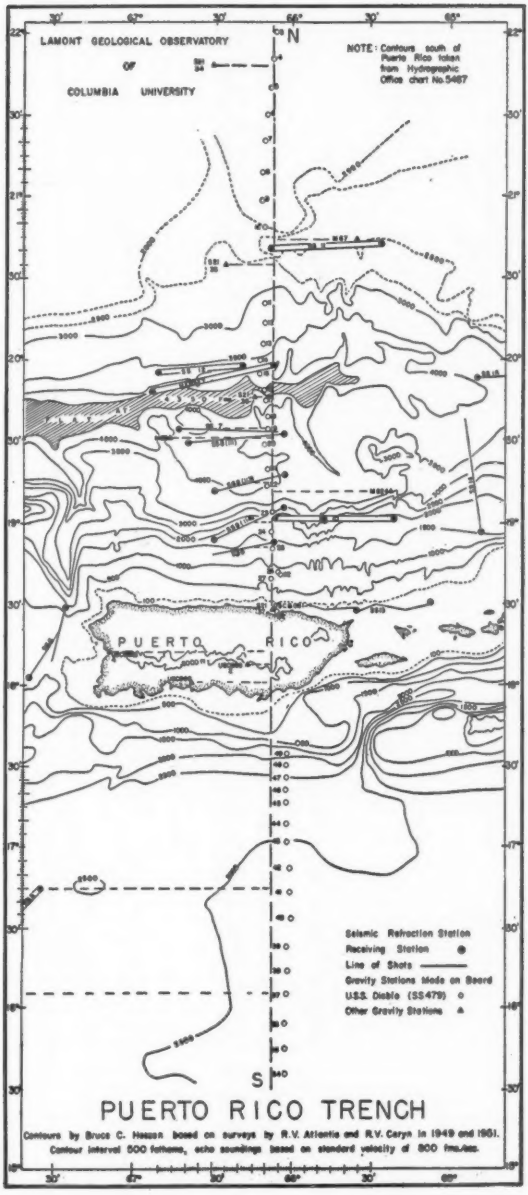
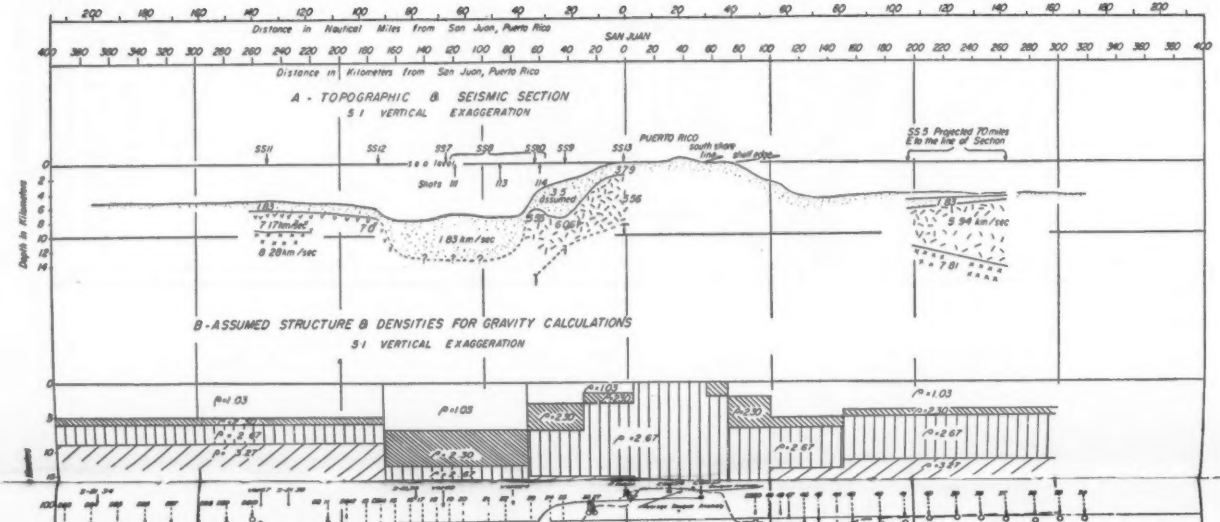

LAMONT GEOLOGICAL OBSERVATORY
OF
COLUMBIA UNIVERSITY
NOTE: Contours south of Puerto Rico taken from Hydrographic Office chart No. 5487
N
S
PUERTO RICO
Seismic Refraction Station
Receiving Station
Line of Shots
Gravity Stations Made on Board
U.S.S. Diablo (SS 479)
Other Gravity Stations
PUERTO RICO TRENCH
Contours by Bruce C. Heezen based on surveys by R.V. Atlantis and R.V. Caryn in 1949 and 1951.
Contour interval 500 fathoms, echo soundings based on standard velocity of 800 fms./sec.

FIGURE 1.—PUERTO RICO TRENCH


Distance in Nautical Miles from San Juan, Puerto Rico
SAN JUAN
Distance in Kilometers from San Juan, Puerto Rico
A - TOPOGRAPHIC & SEISMIC SECTION
5:1 VERTICAL EXAGGERATION
PUERTO RICO
south shore line
shelf edge
SS11
SS12
SS7
SS8
SS10
SS9
SS13
SS5 Projected 70 miles E to the line of Section
Depth in Kilometers
Shots till
3.5 assumed
5.94 km/sec
6.83 km/sec
6.28 km/sec
B - ASSUMED STRUCTURE & DENSITIES FOR GRAVITY CALCULATIONS
5:1 VERTICAL EXAGGERATION
ρ=1.03
ρ=2.67
ρ=2.87
ρ=3.30

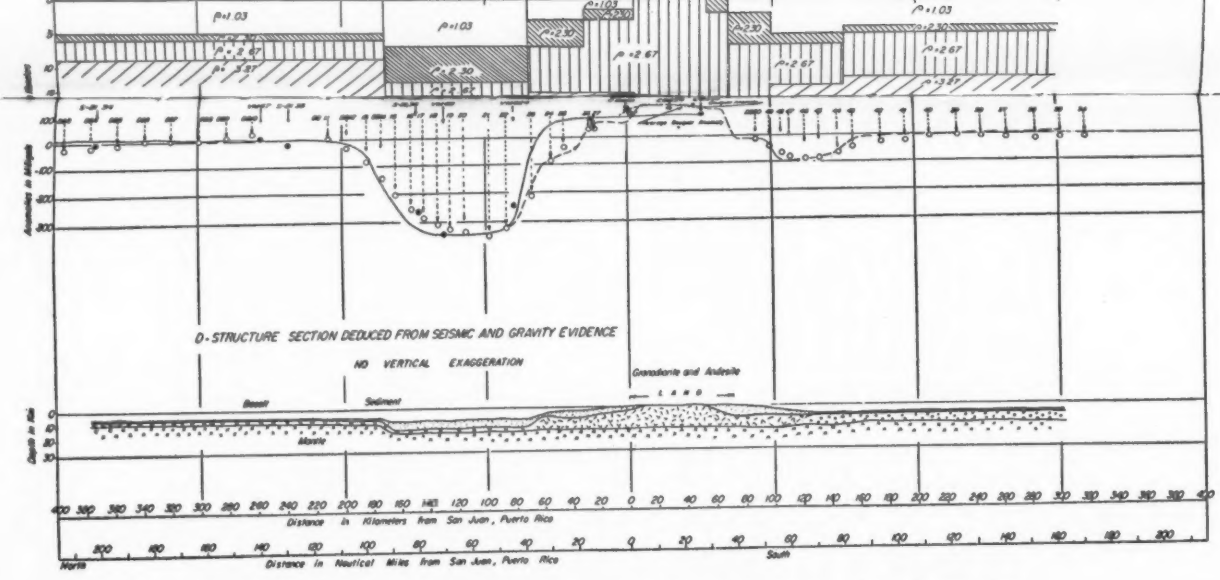


FIGURE 2.—NORTH-SOUTH STRUCTURE AND GRAVITY SECTIONS THROUGH SAN JUAN, PUERTO RICO

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 65, PP. 201-222, 4 PLS. 11 FIGS, MARCH 1964

GEOLOGIC PROBLEMS IN THE SOUTHEAST MISSOURI LEAD DISTRICT

BY E. L. OHLE AND J. S. BROWN, *Editors*[1]

ABSTRACT

This paper is essentially a progress report on the geological study being made in the Southeast Missouri District by the St. Joseph Lead Company. The problems encountered are described and the progress toward their solution is assessed. These problems have consisted of the subdivision of the Bonneterre formation into units that are mappable underground, the recognition of the types of rock alteration accompanying or preceding ore deposition, the recognition of the structural controls of the ore trends, and the collection of data that may indicate the method of ore genesis.

The Bonneterre formation is divided into eight principal zones which are numbered from top to bottom. Contacts between zones are mapped, and structural features are indicated by contouring. Ore in each zone has distinctive characteristics.

Buried Precambrian igneous knobs have localized some ore bodies in the Lead Belt area, but sedimentary arch structures of depositional origin are the outstanding features controlling ore trends. Fracture zones exerted a secondary influence where arches are present and, where arches are absent, were, in some places the dominant ore-controlling structures.

Major alteration effects are dolomitization of limestone, "fingering" and spotting, recrystallization, and removal by solution. Glauconite and adularia have been introduced or redistributed.

Lateral migration of the ore solution was important, and the vertical component of motion almost surely was up rather than down. The marginal position of sphalerite in the galena ore bodies is in accord with the relative solubilities of the two sulfides.

CONTENTS

TEXT

	Page
Introduction	202
Acknowledgments	202
The problems	202
1. Subdivision of the Bonneterre formation	202
2. Rock alteration	207
General discussion	207
Dolomitization	207
Spotting and "fingering"	208
Recrystallization	210
Solution effects	210
Glauconite and adularia	211
3. Structural controls of the ore bodies	211
Regional structure	211
Buried knobs and ridges	212
Sedimentary depositional arches	212
Fracture zones	215
4. Ore genesis	219
Summary	220
References cited	221

ILLUSTRATIONS

Figure	Page
1.—Map of the Lead Belt of southeast Missouri	203
2.—General geologic section in the Lead Belt area	204
3.—Generalized stratigraphic subdivision of the Bonneterre formation	205
4.—Map and section of a typical granite knob structure	213
5.—Map of the mine workings in the Lead Belt	214
6.—Desloge Mine sandy transition contour map	216
7.—Desloge Mine sandy transition isopach map	217
8.—Typical cross section of sand ridge structures	218
9.—Map and section showing progressive southward shift of ridge axes	218
10.—Map of a 7/10 arch	219
11.—Map of a portion of No. 9 Federal Mine	220

Plate	Following page
1.—Gray spotted and "fingered" rock	208
2.—"Fingered" rock in vertical and horizontal section	
3.—"Fan" and "hourglass" structures	
4.—"Roll" structure in No. 9 Federal Mine	

[1] Principal contributions by J. S. Brown, W. S. Detrick, J. A. Emery, J. E. Jewell, J. E. Kiser, E. L. Ohle, F. G. Snyder, R. E. Wagner, W. W. Weigel, and F. B. Whiting. Written and edited by E. L. Ohle and J. S. Brown.

Introduction

For nearly 100 years the great disseminated lead district of Southeast Missouri has supplied 25 to 50 per cent of the primary lead mined annually in the United States. Its total output through 1951 has exceeded 8,000,000 tons of pig lead valued at more than $1,000,000,000. The producing area is nearly 60 miles long by 6 miles wide and is known locally as the Lead Belt. (Fig. 1). More than 90 per cent of the production has come from a triangle with sides some 7 miles in length embracing the towns of Bonne Terre, Desloge, Leadwood, and Flat River.

Despite its great productivity, this area has not until recently received the systematic geological study accorded most other mining districts as accepted operating routine. Broad-scale studies, mostly unpublished, have been made from time to time by the various companies active in the area and by independent students of ore deposition, but all of them either concentrated on specific problems or were handicapped by having limited access to the mines. These limitations are reflected in published articles on the district, most of which emphasize one or another idea as to the origin of the deposits rather than the details of stratigraphic and structural control. The standard reference on the district is still the 1908 report of Buckley, which was written when the great Flat River and Leadwood ore bodies were just undergoing development.

The lack of applied geology in Southeast Missouri is due largely to the ease of exploration by diamond drilling, for ore reserves could be maintained simply by drilling more holes in known mineralized areas. Probably more diamond drilling has been done in the Lead Belt than in any other area of comparable size in the world.

In 1947 the St. Joseph Lead Company instituted an intensive geological study of its properties under the direction of John S. Brown, Chief Geologist, who was transferred at that time from other operations. Since then, from 5 to 10 full-time geologists and several geological-engineering assistants have been engaged in underground mapping and the correlation of underground and drill-hole data. This initial phase of the program is about 75 per cent completed, and this paper is essentially an interim report describing the problems encountered and indicating the degree of progress toward their solution.

Acknowledgments

The authors are indebted to the St. Joseph Lead Company for permission to publish this report.

The Problems

1. *Subdivision of the Bonneterre[2] Formation*

The bulk of the lead ore in Southeast Missouri occurs in the Upper Cambrian Bonneterre dolomite. Its general relations are indicated in Figure 2. Much of the work of subdividing the Bonneterre had been completed by R. E. Wagner and J. E. Jewell prior to 1947 as an accompaniment of the extensive diamond-drilling program. Through a careful study of the cores from thousands of holes, starting about 1936, the rock types were classified, and the formation was divided into zones. Subsequent underground observation has clarified the relationships between certain rock types and has revealed lateral gradations which could not be detected in cores from vertical holes spaced 100 feet or more apart, but the essential features of the stratigraphic column remain unchanged.

The subdivision of the Bonneterre has been complicated by variations in the thickness and appearance of the rock types in the lower half of the formation as originally deposited, and by alteration effects which, in many places, partially or completely obliterate the original character of the beds. Although the total thickness of the formation is 375–400 feet nearly everywhere in the main productive area, lensing and some interlayering of rock types is characteristic of the lower 200 feet. Thinning of one zone is compensated by thickening of other zones to maintain the uniform total. In the upper half of the Bonneterre the sequence and thickness of rock types is much more uniform, and the bedding is more regular.

[2] The spelling Bonneterre refers to the Upper Cambrian rock formation, while Bonne Terre refers to the town of Bonne Terre, Missouri.

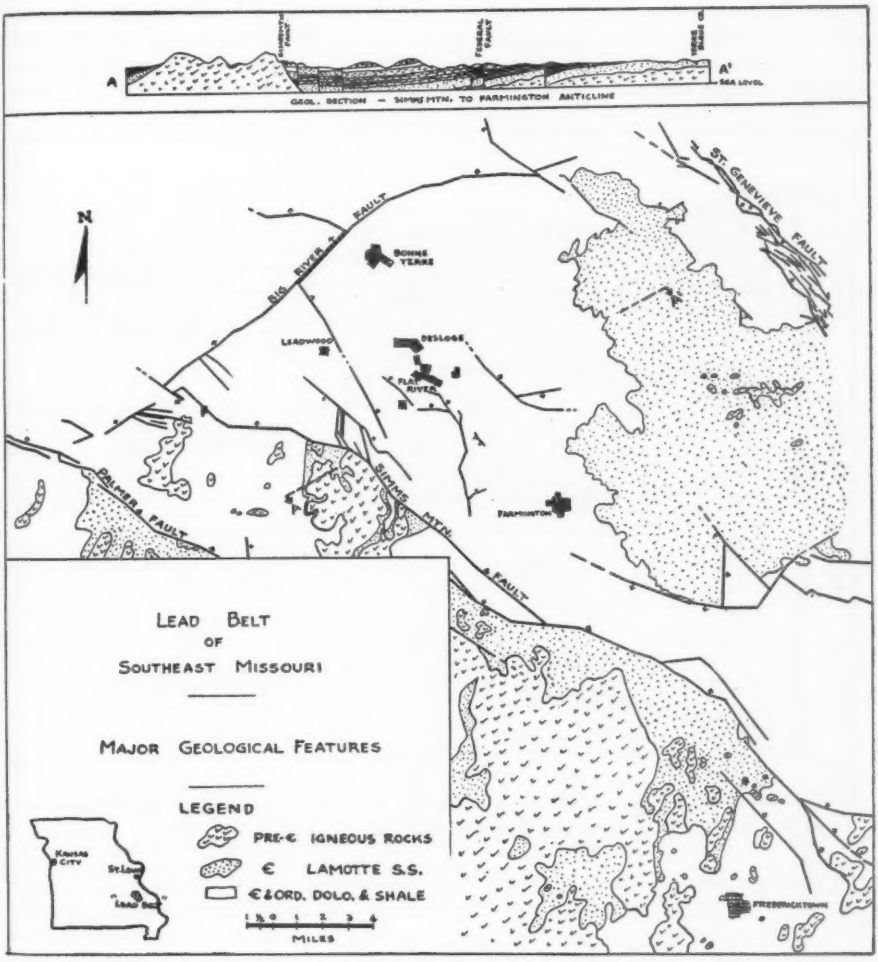


FIGURE 1.—MAP OF THE LEAD BELT OF SOUTHEAST MISSOURI

It is not always possible to determine whether a given rock type is an alteration product or a feature of original sedimentation. Complete agreement is yet to be reached as to the origin of several important types. Certain ones, such as the coarse-grained, glauconitic, light-gray rock near the bottom of the section, seem to have a dual origin, being in some places original and elsewhere secondary.

The formation is divided into eight principal units numbered from top to bottom, 1, 3, 5, 7, 10, 12, 15, and 19 (Fig. 3). The 1-bed is just below the overlying Davis formation, and the 19-bed is the sandy dolomitic phase immediately overlying the Lamotte sandstone. All these units usually are identifiable in ore-bearing areas. However, the distinctive features of several of the zones seem to result from rock alteration—possibly in part hydrothermal and associated with ore deposition. At any rate, outside the general mineralized areas these particular units have not been identified. Several of the units vary in thickness and to some extent in lithology from place to place

Distance in Nautical Miles from San Juan, Puerto Rico
Distance in Kilometers from San Juan, Puerto Rico
SAN JUAN
PUERTO RICO
A - TOPOGRAPHIC & SEISMIC SECTION
5:1 VERTICAL EXAGGERATION
B - ASSUMED STRUCTURE & DENSITIES FOR GRAVITY CALCULATIONS
5:1 VERTICAL EXAGGERATION

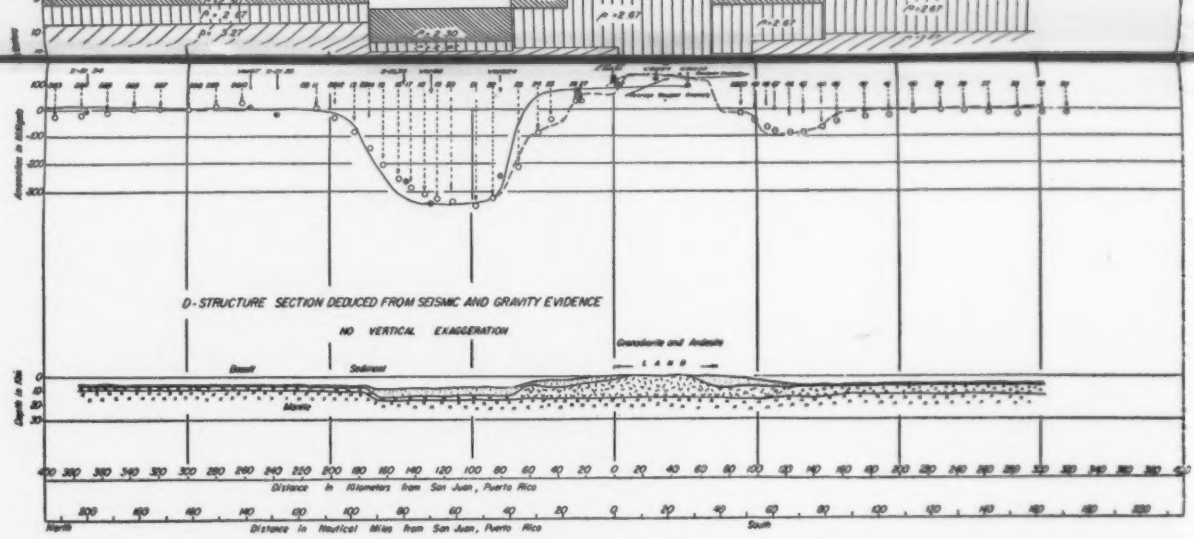


FIGURE 2.—NORTH-SOUTH STRUCTURE AND GRAVITY SECTIONS THROUGH SAN JUAN, PUERTO RICO